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Boris S. Sokolov

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Sm-Nd Isotopic Systematics of the Kurumkan Subgroup, the Iengra Group of Aldan Shield

V. P. Kovach*, A. B. Kotov*, E. B. Sal'nikova*, V. I. Berezkin, A. P. Smelov**, and N. Yu. Zagornaya***

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Abstract—The first data on the Nd isotopic composition in high-Al gneisses and associated garnet–hypersthene–biotite plagiogneisses of the Kurumkan Subgroup, the Iengra Group of Aldan Shield are presented. The obtained results show that rocks of the subgroup accumulated after ~2.9 Ga, i.e., during the Late Archean or Early Proterozoic, and not the Early Archean evolutionary stage of the shield. Protoliths of garnet–hypersthene–biotite gneisses were not older than 2.3 Ga, and their formation marks the Early Proterozoic stage of continental crust generation in the Aldan granulite–gneiss region.

Key words: *Sm-Nd isotopic systematics, high-Al gneisses, Kurumkan Subgroup, metamorphism, Late Archean, Early Proterozoic, Aldan Shield.*

INTRODUCTION

After the pioneering works of Korzhinsky (1936, 1939), the Iengra Group of the Aldan Shield, comprising a characteristic association of quartzites and high-Al gneisses, was traditionally considered as a typical example of the high-grade metamorphic supracrustal complexes of the Katarchean (Cherkasov, 1979; Dzevanovskii, 1958; *Dokembriiskaya ...*, 1988; Ez, 1983; Frumkin, 1981; Neelov and Milkevich, 1979; *Precambrian ...*, 1993; *Rannii ...*, 1986; Reutov, 1972; Rudnik, 1975; Salop and Travin, 1974), and this concept was used for interregional and intercontinental correlations (*Dokembriiskaya ...*, 1988; *Precambrian ...*, 1993; Salop, 1983). However, convincing arguments favoring the Katarchean age of this group are absent (Tugarinov *et al.*, 1967, 1976), and the concept is based on the high grade of rock metamorphism (granulite facies), along with some geological and structural data. Recent data on geochronology and isotopic geochemistry showed that protoliths of tonalite–trondhjemite gneisses from the presumably basement complex, and metavolcanics of the Fedorov Group in the eastern part of the Nimnyr block of the Aldan granulite–gneiss terrane (AGGT) are of the Early Proterozoic age (Kovach *et al.*, 1992; Kovach, 1994; Kotov *et al.*, 1995; Sal'nikova *et al.*, 1992; Kovach *et al.*, in press). They also demonstrated that in the northwestern part of the block, the Early Proterozoic reworking strongly affected the Archean infracrustal gneisses (Nutman *et al.*, 1992; Sal'nikova, 1993). In addition, it was noted that Kurumkan rock associations are similar to both the Early Archean (*Drevneishie ...*, 1989) and Early Proterozoic (Nutman *et al.*, 1992) supracrustal complexes. Thus, the Iengra Group geochronology in the Aldan Shield is a topical problem.

In this paper, this problem is considered in light of new Sm-Nd isotopic data for clastic metasedimentary rocks. As the Sm/Nd ratios remain almost constant during weathering, transportation, diagenesis, and metamorphism (De Paolo, 1988; Taylor and McLennan, 1975), the study of the Nd isotope composition in these rocks provides an opportunity to measure the average model age of their provenance (McCulloch and Wasserburg, 1978; O'Nions *et al.*, 1983), and to evaluate the maximum age limit of supracrustal complexes (Miller *et al.*, 1986).

GEOLOGICAL REVIEW

At the rank of an individual unit of the Iengra Group, the Kurumkan Formation of the Aldan Shield was first distinguished by Salop and Travin (1974). It is correlated with the lower portion of the Upper Aldan Formation of the Iengra Group by Dzevanovskii (1958), Frumkin (1981), Reutov (1972), and others. In addition, Salop and Travin discriminated within the group the Ayanakh and Suon-Tit formations. Later, Dook and his colleagues (*Dokembriiskaya ...*, 1988; *Rannii ...*, 1986; *Precambrian ...*, 1993) united all of these formations under the name of the Kurumkan Subgroup, with the exception of their enderbite–, charnokite–, and granite–gneiss associations (the Nimnyr Formation of the Upper Aldan Group) attributed to the nonstratified

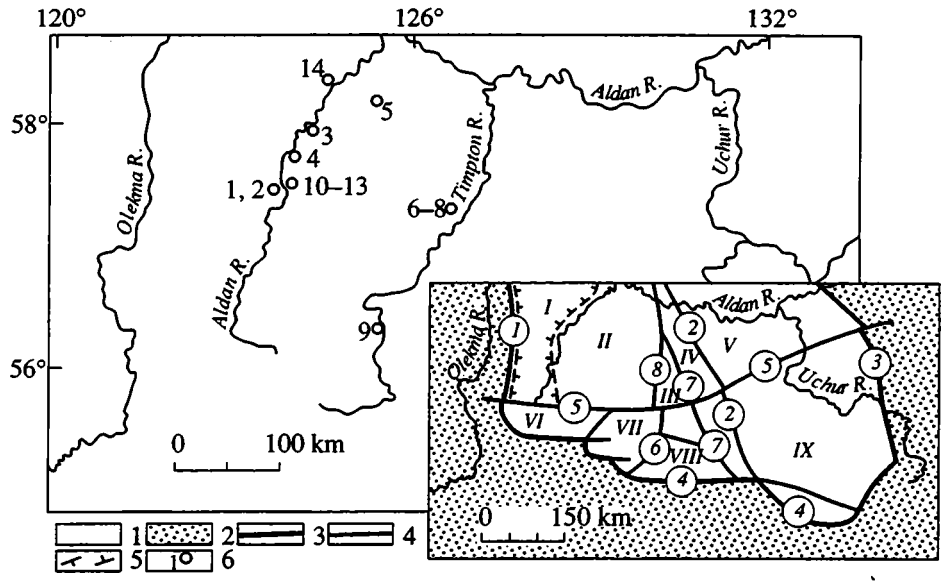

Fig. 1. Sites of samples used for Sm-Nd isotopic measurements; the AGGT structural zoning is after Neelov and Milkevich (1979) with our modifications.

(1) Aldan granulite–gneiss terrane; (2) Early Precambrian structures surrounding the Aldan Shield; (3) boundary fault zones of the massif; (4) block-separating faults; (5) inferred boundaries of the junction zone between the Olekma granite–greenstone and Aldan granulite–gneiss terranes; (6) sampling sites (numbers as in Table 2). (*I*) Junction zone between the Olekma granite–greenstone and Aldan granulite–gneiss terranes; blocks: (*II*) Nimnyr; (*III*) Seim; (*IV*) Kholbolokh; (*V*) Sunnagin; (*VI*) Zverev; (*VII*) Melemken; (*VIII*) Sutam; (*IX*) Tyrkan. Numbers in circles for fracture zones: (*1*) Amga; (*2*) Tyrkanda; (*3*) Ulkan; (*4*) South Aldan; (*5*) Tungurcha–Gynym; (*6*) Gonam; (*7*) Nuyam; (*8*) Timpton thrust.

infracrustal complex. Detailed characteristics of the geological structure, lithology, and chemical composition of the Kurumkan rocks are presented in several publications (*Dokembriiskaya ...*, 1988; *Drevneishie ...*, 1989; Neelov and Milkevich, 1979; *Rannii ...*, 1986; Travin, 1977; *Precambrian ...*, 1993).

Rocks of the Kurumkan Subgroup are widespread in the Nimnyr and Melemken blocks of western AGGT (Fig. 1), where they are exposed as fragments of complexly folded structures up to 15 × 30 km in dimension, which are surrounded by enderbite– and granite–gneisses of the infracrustal complex in the deeply eroded part of the Nizhnii Timpton dome (*Dokembriiskaya ...*, 1988; *Rannii ...*, 1986; *Precambrian ...*, 1993). In the western part of the Nimnyr block, the Kurumkan Subgroup includes a characteristic association of quartzites and high-Al gneisses corresponding in chemical composition to ultrasilicites, aleuropelites, and pelites (*Dokembriiskaya ...*, 1988; *Rannii ...*, 1986; *Precambrian ...*, 1993). Its section in the eastern part of the block is less diverse, and comprises only high-Al gneisses (pelites and hydromica subsialites). The Kurumkan succession of the Melemken block consists of high-Al gneisses and quartzites serially changing in composition from ultrasilicites to hydromica subsialites, and enclosing lenses of calc-siliceous rocks, calciphyres, ferruginous and diopside quartzites. In the northwestern part of the Nimnyr block, the U–Pb zircon age for tonalite–trondhjemite gneisses of the infracrustal complex is 33355 ± 3 Ma (SRIMP; Nutman *et al.*, 1990, 1992) or ~3.6 Ga (Morozova *et al.*, 1989). The model $T_{Nd}(DM)$ ages for these gneisses are in the range of 3.5–3.8 Ga (Kotov *et al.*, 1993; Sal'nikova *et al.*, 1992; Sal'nikova, 1993). In the eastern part of the Nimnyr block, the concordant age value of zircons from the tonalite–trondhjemite gneisses of the infracrustal complex is 2011 ± 3 Ma, while their Sm-Nd model age is ~2.4 Ga (Kovach, 1994; Kotov *et al.*, 1995).

High-Al rocks of the Kurumkan Subgroup are represented by garnet–cordierite–biotite, garnet–sillimanite–cordierite–biotite, sillimanite–biotite, and cordierite–sillimanite–biotite two-feldspar gneisses. They are of the Na–K-type, and show low Ca and high Fe contents, and a high degree of iron oxidation (Table 1). Data available for the eastern part of the Nimnyr block suggest that high-Al gneisses of the Kurumkan Subgroup and biotite–hypersthene plagiogneisses of the infracrustal complex cannot be attributed to a single differentiation trend (*Dokembriiskaya ...*, 1988; *Rannii ...*, 1986; *Precambrian ...*, 1993).

Quartzites of the Kurumkan Subgroup are mostly of monomineral composition; rarer varieties are represented by feldspar quartzites grading into quartzite–gneisses, and by biotite, sillimanite, and magnetite quartzites. According to Travin (1977), quartzites of this subgroup could have been originally represented by both the chemogenic and terrigenous sediments. The Pb–Pb zircon age of the Kurumkan quartzites is 2.2–3.0 Ga (*Dokembriiskaya ...*, 1988; *Drevneishie ...*, 1989; Tugarinov *et al.*, 1967; *Precambrian ...*, 1993).

Table 1. Chemical composition (wt %) of representative rocks of the Kurumkan Subgroup, the Iengra Group of the Aldan Shield

Sample	B-1906/10	B-1906/21	B-1907/6	B-1907/6a	B-1907/17	B-1907/18a	B-1922/1	B-2893	V-31v
SiO_2	53.40	57.27	75.06	63.27	62.58	62.02	66.70	54.31	58.47
TiO_2	1.05	1.10	0.57	0.88	0.71	0.73	0.78	1.01	1.14
Al_2O_3	25.40	22.03	10.49	15.06	14.50	14.62	15.39	21.67	20.61
Fe_2O_3	5.17	0.96	0.60	0.83	0.65	0.47	0.01	3.13	1.05
FeO	5.03	10.06	4.38	5.52	6.35	7.34	6.16	6.75	10.50
MnO	0.08	0.10	0.06	0.07	0.10	0.10	0.03	0.10	0.07
MgO	1.66	2.45	1.80	4.22	4.21	4.33	2.02	2.70	2.60
CaO	0.45	1.33	1.51	3.88	4.20	3.95	1.84	1.64	1.34
Na_2O	1.68	1.43	2.40	2.73	3.01	2.90	5.03	3.26	1.03
K_2O	5.31	2.16	1.85	1.79	2.19	2.09	1.76	4.06	2.54
P_2O_5	0.27	0.14	0.10	0.21	0.21	0.18	0.06	0.11	0.08
LOI	1.06	1.17	0.40	0.99	1.06	1.00	0.28	1.39	0.59
Total	100.56	100.20	99.22	99.45	99.77	99.73	100.06	100.13	100.02

Note: Samples B-1906/10, B-1906/21, B-2893, and V-31v are high-Al gneisses; other samples correspond to garnet–hypersthene–biotite gneisses.

In addition to quartzites and high-Al gneisses, the Kurumkan Subgroup includes concordant bodies of two-pyroxene (±amphibole, ±biotite) and amphibole schists associated with amphibolites, all of which correspond in composition to tholeiitic basalts (Velikoslavinsky *et al.*, 1990, 1993; *Dokembriiskaya ...*, 1988; *Rannii ...*, 1986; *Precambrian ...*, 1993). Boundaries of basic schist areas are discordant relative to contacts between the various rock types of the Kurumkan Subgroup, and between the latter and infracrustal complex (*Rannii ...*, 1986). Because geochemical relations between high-Al gneisses and quartzites are lacking, Travin (1977) concluded that these schists are embedded into the Kurumkan succession like trapps and/or dykelike bodies. In addition, Velikoslavinsky *et al.* (1990, 1993) showed that Kurumkan basic schists are associated with metavolcanics of intermediate and acid composition, typical of the andesite–trachyte and dacite–rhyolite series, which transformed into enderbite and charnokite gneisses.

In the stratotype locality near Kurumkan (the Aldan River upper courses), quartzites, high-Al gneisses, basic schists, and amphibolites alternate with hypersthene–biotite, garnet–hypersthene–biotite, and garnet–biotite plagioschists and gneisses[1] (Kitsul, 1989) with chemical compositions similar to graywackes, tuffites, and magmatic rocks of intermediate to acid composition (Table 1). Initially attributed to the Kurumkan succession of the Upper Aldan Group (Dzevanevskii, 1958; Neelov and Milkevich, 1979; Salop and Travin, 1974; Frumkin, 1981), these rocks were excluded from it by Dook and his colleagues (*Rannii ...*, 1986) on the basis of petrochemical investigations.

According to present concepts (*Dokembriiskaya ...*, 1988; *Rannii ...*, 1986; Travin, 1977; *Precambrian ...*, 1993), the Kurumkan rock association of the Iengra Group was accumulated in a cratonic setting lacking manifestations of volcanism. In the opinion of Neelov and Milkevich (1979), this association in the western part of the Nimnyr block represents continental deposits grading eastward into lagoonal and marine geosynclinal facies.

In addition to the Kurumkan Subgroup and infracrustal complex, the studied region also comprises widespread, mostly volcanogenic rocks of the Fedorov Group, which occur in the allochthonous position (*Dokembriiskaya ...*, 1988; *Rannii ...*, 1986; *Precambrian ...*, 1993). In the eastern part of the Nimnyr block, $T_{Nd}(DM)$ values for the intermediate and acid metavolcanics of the Fedorov Group are ~2.2–2.3 Ga (Kovach, 1994; Kovach *et al.*, in press), which is close to the Pb–Pb isochron age (2000 ± 80 Ma) of these rocks with the mantle values of $\mu_1 = 8.02 \pm 0.08$ (Neymark, 1981).

In the Nimnyr block of AGGT, the nonstratified infracrustal complex and rocks of the Kurumkan and Fedorov units experienced multistage structural and metamorphic transformations under conditions of granulite facies (*Dokembriiskaya ...*, 1988; *Rannii ...*, 1986; *Precambrian ...*, 1993); this was partly related to the formation of deep-seated viscous thrusts of different ages (Kotov *et al.*, 1995; Sal'nikova, 1993). It is well established at present that in the northwestern part of AGGT, rocks of the Kurumkan Subgroup were involved in thrusting processes whose maximum age limit is about 2.4 Ga (Sal'nikova, 1993). The time-range of the most intense tectonic–magmatic reworking of infracrustal

[1] Hereafter, all of these rocks are referred to as garnet–hypersthene–biotite gneisses.

gneisses in this area of AGGT is estimated to be 1907–1962 Ma (Kotov *et al.*, 1993; Nutman *et al.*, 1992). In eastern zones of the Nimnyr block, the earliest structural transformations of high-Al gneisses of the Kurumkan Subgroup and tonalite–trondhjemite gneisses of the supposed basement complex are not older than 2011 ± 3 Ma, whereas the tectonic juxtaposition of these and Fedorov Group rocks, and culmination of granulite metamorphism occurred not earlier than 1992 ± 1 Ma ago (Kovach, 1994; Kotov *et al.*, 1995).

ANALYTICAL PROCEDURE

The Sm and Nd isotope composition was measured at the Institute of Precambrian Geology and Geochronology, Russian Academy of Sciences, by the procedures described earlier (Kotov, *et al.* 1995; Neymark *et al.*, 1993). The total blank during measurements was 0.03–0.2 ng for Sm and 0.1–0.5 ng for Nd. Measured $^{143}Nd/^{144}Nd$ ratios were normalized to $^{146}Nd/^{144}Nd = 0.7219$ and corrected relative to $^{143}Nd/^{144}Nd = 0.511860$, accepted for the La Jolla Nd standard. The analytical precision is ±0.5% (2σ) for Sm and Nd concentrations; for $^{147}Sm/^{144}Nd$ and $^{143}Nd/^{144}Nd$ ratios, it is ±0.5 and ±0.005%, respectively. For 11 measurements in the La Jolla Nd standard, the weighted-mean $^{143}Nd/^{144}Nd$ ratio is 0.511894 ± 8 (2σ). Calculating the $\varepsilon_{Nd}(0)$ values and model $T_{Nd}(DM)$ ages, we used isotopic parameters of the chondrite undifferentiated reservoir (CHUR) after Jacobsen and Wasserburg (1984), and depleted mantle (DM) after Goldstein and Jacobsen (1988).

RESULTS AND DISCUSSION

The Sm-Nd isotopic parameters were measured in high-Al gneisses, which were sampled from various rock associations in different distribution areas of the Kurumkan Subgroup, and in garnet–hypersthene–biotite gneisses of the type section. The sampling localities are shown in Fig. 1.

$T_{Nd}(DM)$ values for the garnet–hypersthene–biotite gneisses of the Kurumkan Subgroup range from ~2.3 to 2.4 Ga (Table 2, Fig. 2) and suggest the Early Proterozoic age of the provenance, if the gneisses are formed after the graywacke protoliths. If their protoliths were tuffites, then the model ages obtained may indicate that an old crustal material of source areas was mixed with a "mantle" material, whose age is < 2.3 Ga. In the case of the magmatic origin of protoliths, the model Sm-Nd ages of these gneisses should imply the Early Proterozoic age of magma-generating substratum.

The oldest $T_{Nd}(DM)$ value of ≈3.5 Ga is calculated for Sample Dr-0316/12 of the cordierite–garnet–biotite gneiss (Table 2, Fig. 2), which was collected in the enderbite gneisses locality (Fig. 1) of the Early Archean infracrustal complex (Morozova *et al.*, 1989; Nutman *et al.*, 1990, 1992). Other studied samples of high-Al gneisses yielded the model age values ranging within the narrow interval of ~2.9–3.1 Ga (Table 2, Fig. 2) despite the different character and geographic position and composition of the sampled Kurumkan sections.

The results of our Sm-Nd isotopic investigations show that the Kurumkan Subgroup is not older than ~2.9 Ga, and should be referred not to the Early, but to the Late Archean, or even younger evolutionary stage of the Aldan Shield. Protoliths of the studied garnet–hypersthene–biotite gneisses were represented by rocks not older than 2.3 Ga, whose formation corresponded to the Early Proterozoic epoch of continental crust generation within the AGGT.

The clear difference between the Sm-Nd model ages of the high-Al and garnet–hypersthene–biotite gneisses (Table 2, Figs. 2 and 3) may imply that their formation was discrete in time and/or space. The chemical composition of the garnet–hypersthene–biotite gneisses (suggesting their similarity to graywackes, tuffites, and andesite–dacites), along with the comparable model ages which were obtained for these rocks, tonalite–trondhjemite gneisses of the infracrustal complex, and intermediate to acid metavolcanics of the Fedorov Group from the eastern part of the Nimnyr block (Fig. 3), suggest that protoliths of the former rock type could be genetically related to the latter rock complexes. In this case, we may assume that the tectonic juxtaposition of the high-Al and garnet–hypersthene–biotite gneisses occurred during the formation period of the Early Proterozoic thrust faults widespread within the AGGT (Kovach, 1994; Kotov *et al.*, 1995; Sal'nikova, 1993). At the same time, it is unclear as to how the studied garnet–hypersthene–biotite rocks are related to the intermediate–acid metavolcanics of the Kurumkan(?) succession distinguished by Velikoslavinsky *et al.* (1990, 1993).

The infracrustal tonalite–trondhjemite gneisses in both the northwestern and eastern parts of the Nimnyr block differ in model Sm-Nd ages from most of the studied samples of high-Al gneisses of the Kurumkan Subgroup (Fig. 3). This implies that the infracrustal complex was not the only protolith of the high-Al gneisses.

A possible formation model of the latter rock group is as follows. In the western part of the Nimnyr block of the AGGT, rocks of the Kurumkan Subgroup accumulated on the Early Archean continental crust [$T_{Nd}(DM) \approx 3.5–3.8$ Ga]. In the eastern part of the block, protoliths of the high-Al gneisses were represented by lagoonal and marine sediments whose provenance was located in a continental area with exposures of the Early Archean rocks and Early Proterozoic volcanics of the Fedorov Group formed under island-arc conditions. Concurrently, or soon after the formation of these protoliths in an active continental margin, rocks of the tonalite–trondhjemite composition were formed, which are 2011 ± 3 Ma old (Kovach, 1994; Kotov *et al.*, 1995) and belong, according to Dook and his colleagues (*Dokembriiskaya* ..., 1986; *Rannii* ..., 1986;

Precambrian ..., 1993), to the nonstratified infracrustal complex. Because their $\varepsilon_{Nd}(T)$ values approach zero and $\delta^{18}O$ values are about 11‰, these rocks may represent the crystallization products of parental magmas that assimilated the original material of high-Al gneisses (Kovach, 1994). Later on, the protoliths of high-Al gneisses were thrust over rocks of the infracrustal complex, and both rock units experienced structural and metamorphic transformations related to the initial collision between the island arc and continent. During the period preceding 1992 ± 1 Ma, rocks of the Kurumkan and infracrustal complexes were tectonically grouped together with volcanics of the Fedorov Group, and all suffered the next deformation and metamorphism event. In this case, we may explain why model ages of the high-Al gneisses do not depend on their geographic localization, and why the Kurumkan Subgroup includes rocks with both the Early Archean and Early Proterozoic model ages. According to this model, at least a part of the high-Al gneisses of the Kurumkan Subgroup was formed approximately 2.0–2.2 Ga ago, i.e., during the Early Proterozoic evolutionary stage of the Aldan Shield.

It is also possible that protoliths of the high-Al gneisses with model $T_{Nd}(DM)$ ages of about 2.9–3.1 Ga have been derived from the Late Archean rocks, which are either still unknown, or entirely eroded in the western part of the AGGT. The Late Archean stage of crust generation is well manifested in the eastern part of the Olekma granite–greenstone area (Sal'nikova, 1993; Sal'nikova *et al.*, 1992), west of the studied region (Fig. 1). For the AGGT, this stage of geologic evolution can be indirectly argued by the values of the Sm-Nd two-stage model ages [$T_{Nd}(DM\text{-}2st)$] of some Early Proterozoic (~1.97–1.92 Ga old) granitoids, which are ~3.0–3.1

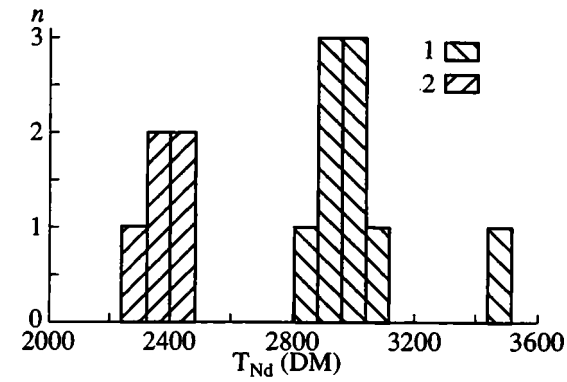

Fig. 2. Histogram illustrating difference between model $T_{Nd}(DM)$ ages of high-Al (1) and garnet–hypersthene–biotite (2) gneisses of the Kurumkan Subgroup, the Iengra Group of the Aldan Shield (*n*—number of measurements).

Table 2. Sm-Nd isotopic data for rocks of the Kurumkan Subgroup, the Iengra Group of the Aldan Shield

No.	Sample no.	Sm, μg/g	Nd, μg/g	$^{147}Sm/^{144}Nd$	$^{143}Nd/^{144}Nd$ ($\pm 2\sigma_{av.}$)	$\varepsilon_{Nd}(0)$	$T_{Nd}(DM)$
1	B-1906/10	8.04	46.18	0.10563	0.511030 ± 8	−31.4	2975
2	B-1906/21	8.31	49.06	0.10270	0.510944 ± 9	−33.0	3013
3	Dr-0316/12	3.25	17.04	0.11564	0.510912 ± 9	−33.7	3456
4	Dr-0261v	8.32	47.68	0.10579	0.511029 ± 16	−31.4	2980
5	B-2893	8.01	50.62	0.09605	0.510912 ± 9	−33.7	2885
6	N-1/1	9.33	57.53	0.09835	0.510978 ± 8	−32.4	2856
7	N-4/2	6.44	36.51	0.10696	0.511106 ± 8	−29.9	2904
8	K-188	6.37	37.60	0.10271	0.511009 ± 10	−31.8	2925
9	V-31b	4.53	22.66	0.12134	0.511286 ± 10	−26.4	3060
10	B-1907/b	7.33	41.53	0.10710	0.511429 ± 16	−23.6	2453
11	B-1907/6a	4.74	24.72	0.11624	0.511588 ± 10	−20.5	2435
12	B-1907/17	3.49	19.22	0.10968	0.511519 ± 6	−21.8	2383
13	B-1907/18a	3.65	19.83	0.11134	0.511580 ± 4	−20.6	2331
14	B-1922/1	6.79	32.58	0.11054	0.511618 ± 9	−19.9	2258

Note: The rock mineral composition and sampling locality: B-1906/10—sillimanite–cordierite gneiss, the Aldan River 2 km upstream of the settlement of Kurumkan; B-1906/21—garnet–sillimanite–cordierite–biotite gneiss, the same locality; Dr-0316/12—cordierite–garnet–biotite gneiss, the right side of the Aldan River 15 km downstream of the Grekov Rapids; Dr-0261v—garnet–cordierite gneiss, the right side of the Aldan River 10 km downstream of the settlement of Suon-Tit; B-2893—cordierite gneiss, the locality near the city of Aldan; N-1/1—cordierite–garnet–hypersthene gneiss from the Kurung-Khoonku–Egete-Timpton interfluve area; N-4/2—sillimanite–cordierite–hornblende gneiss, the same locality; K-188—biotite–garnet–sillimanite–cordierite–hypersthene gneiss, the same locality; V-31b—garnet–sillimanite gneiss, the Timpton River upper courses; B-1907/6—garnet–biotite plagiogneiss, the right side of the Aldan River 2 km upstream of the settlement of Kurumkan; B-1907/6a—garnet–hypersthene–biotite plagiogneiss, the same locality; B-1907/18a—biotite–hypersthene plagiogneiss, the same locality; B-1922/1—garnet–hypersthene–biotite plagiogneiss, the right side of the Aldan River ~50 km downstream of the Chuga mouth.

Letter indices mark samples of V.I. Berezkin (B), V.S. Shkodzinskii (V), G.M. Drugova (Dr), A.B. Kotov (K), and N.Yu. Zagornaya (N).

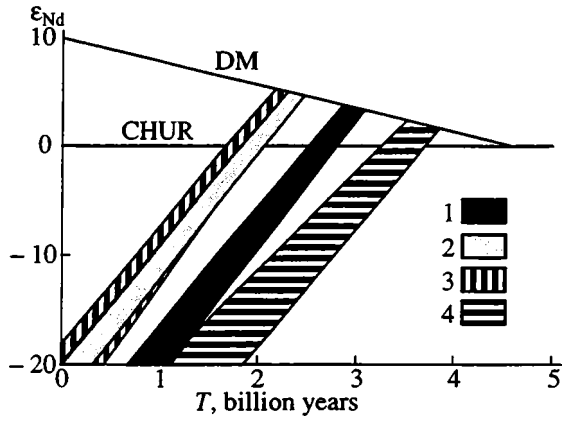

Fig. 3. ε_{Nd} versus T diagram for studied rocks: high-Al (1) and garnet–hypersthene–biotite (2) gneisses of the Kurumkan Subgroup; intermediate to acid metavolcanics of the Fedorov Group and infracrustal tonalite–trondhjemite gneisses (3) from the eastern part of the Nimnyr block; the infrastructure rock types from the northwestern part of the block (4).

and ~2.8–3.0 Ga respectively for the Nimnyr (northwestern and eastern parts) and Seim blocks (Kovach, 1994; Kotov *et al.*, 1994, 1995; Sal'nikova, 1993; Sal'nikova *et al.*, 1992). However, it is unclear at present whether these ages really reflect the Late Archean stage of crust generation in the AGGT, or they simply imply the mixing between material of different age.

CONCLUSION

The results of our Sm-Nd isotopic systematics for high-Al gneisses of the Kurumkan Subgroup (the Iengra Group of the Aldan Shield) and associated garnet–hypersthene–biotite plagiogneisses allow us to state the following:

(1) The model $T_{Nd}(DM)$ ages of the high-Al gneisses (metapelites) are ~2.9–3.1 and 3.5 Ga. They suggest that the accumulation of the Kurumkan Subgroup terminated after ~2.9 Ga, i.e., during the Late Archean or Early Proterozoic evolutionary stage of the Aldan Shield.

(2) The source rocks for protoliths of the associated garnet–hypersthene–biotite gneisses were not older than 2.3 Ga. Their formation marks the Early Proterozoic stage of the continental crust generation in the AGGT.

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Geochronology and Formation Subdivisions of the Lower Precambrian in the Northwestern Ladoga Region

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Abstract—A brief geological review of the northwestern Ladoga region and systematics of the Lower Precambrian subdivisions in this territory are presented. The respective stratigraphic scale and supplementary geologic scheme include the following stratigraphic and metamorphic units applicable in cartography: (1) the 2700-Ma-old North Ladoga granite–gneiss association of the Archean; (2) the Sortavala Group with lower metaclastic and upper metatholeiitic subgroups of the Lower Karelian (Jatulian, Ludicovian), which are 2120–1980 Ma old; and (3) the Ladoga Group of the Upper Karelian (Kalevian) metaturbidites, which are 1910–1880 Ma old. Eight metamorphic suites of the Lahdenpohja Group are referred to the undivided Ludicovian-Kalevian. The Late Karelian plutonic rocks are subdivided into the following units: the Kurkijoki and Lauvatsaari–Impiniemi complexes of the early orogenic stage (1880–1860 Ma) represented by metanorite–enderbite and metadiorite–tonalite associations respectively; the Tervu metadiorite–granite complex of the late orogenic stage (1830–1810 Ma); and the Ojajärvi monzonite-granite association (1800–1760 Ma) of postorogenic origin. Geological relations between these subdivisions in metamorphic zones of the Ladoga region and pertinent geochronological problems are discussed.

Key words: *Ladoga region, Archean, Lower Proterozoic, Ludicovian, Karelian, stratigraphy, magmatism, metamorphism, regional subdivision scheme.*

The northwestern Ladoga region is part of a vast metamorphic belt spread across the southern Baltic Shield. Rocks of this territory can be interpreted as the constituents of a zonal metamorphic complex exposed between the Karelian Massif of the Archean in the northeast and the Wiborg rapakivi pluton in the southwest (Fig. 1). The territory is located in poorly exposed marshy lowland with the exception of the narrow northwestern coastal zone of the Ladoga Lake, where rock outcrops are abundant. This well-studied zone is the key area for understanding the geological structure of the Ladoga complex.

Systematic geological investigations in the northwestern Ladoga region were initiated in the last century. From the very beginning, the most detailed data have been obtained in the northern part of the region, which comprises a separate North Ladoga zone bounded by the Meiyeri fault on the south. In due time, geological data from this zone and other areas adjacent to the Karelian Massif gave rise to Eskola's famous concept of the mantled gneiss domes. In fact, the emergence of the concept was a result of geological studies, which showed that metamorphic formations of the region are represented by two complexes different in age, which were later attributed to the Archean and Lower Proterozoic. The latter was subdivided afterwards into the lower Sortavala Group of mostly volcanogenic origin, and the upper Ladoga Group composed of metaterrigenous rocks. This principal scheme is valid even at present, although the Lower Proterozoic groups are now subdivided into a series of constituent formations.

The metamorphism of rocks in the West Ladoga zone is of a higher grade, and their stratigraphic subdivision is less reliable than that in northern areas. There are two principal viewpoints regarding this problem. One of them suggests that both zones are composed of concurrent rock formations, and therefore, rocks of the West Ladoga zone can be divided by applying the correlation principles to the same Archean and Lower Proterozoic units (including the Sortavala and Ladoga groups) as in the North Ladoga zone (*Geologicheskoe* ..., 1970). Another idea is to consider the high-grade metamorphic formations of the West Ladoga zone as constituting a vast inlier of the Archean basement among folded structures of Svecofennides (Salop, 1979). The diversity of opinions is not surprising, because the stratigraphic classification of structural and rock complexes in this poorly studied zone is unsound and insufficiently constrained by the isotopic dating.

Nevertheless, recently obtained new data on metamorphic rocks of the Ladoga complex (Svetov and Sviridenko, 1992; Shuldiner *et al.*, 1995a, 1995b) afford the opportunity to revise the former concepts. In this paper,

we intend to systematically consider these data in order to unify the distinguishable stratigraphic units. This unification is especially important for the future compilation of a new series of geological maps of the region. Elaborating our stratigraphic scheme (table), we used a new approach to maping and classification of metamorphic complexes. It grants the admittance of specific cartographic units, such as metamorphic suites, when the original stratification or plutonic origin of rocks cannot be confidently inferred (Shuldiner, 1992). In addition to normal stratigraphic and plutonic units, rock formations of this type are widespread in the Ladoga region (Fig. 2). Protoliths for metamorphic rocks of the region are either the Archean, or Early Proterozoic (Karelian) in age, whereas their metamorphism has also been primarily related to the Svecofennian events of the Early Proterozoic time.

It is pertinent to explain how we used the above terms in our paper. Our stratigraphic nomenclature is based on the Precambrian stratigraphic scale accepted in 1990 for the territory of the former Soviet Union and published by Semikhatov *et al.* (1991), where the term Karelian is used as a synonym of the Lower Proterozoic (Paleoproterozoic). In recent publications outside Russia, this term is commonly used to designate only the basal part of the Lower Proterozoic, while younger formations within the interval from the Kalevian level to late orogenic complexes of Svecofennides are attributed to the Svecofennian (Gaal and Gorbatschev, 1987). In this paper, we use the last term in a geotectonic, and not stratigraphic sense (like the term Variscan, etc.) to indicate events of the same age interval.

ARCHEAN ROCKS

The North Ladoga granite-gneiss association. As it was already noted, the Archean comprises ancient basement rocks of the North Ladoga zone, where they are exposed in central areas of mantled gneiss domes. The rocks are represented by dominant medium-grained leuco- and mesocratic biotite, or biotite–amphibole gneisses and migmatites with clear schistosity, which is deformed into narrow, mostly isoclinal folds of different ranks. Subordinate rocks are schists of basic and intermediate composition. It is a common practice to suggest that gneisses and migmatites are the products of metamorphism and remobilization that affected the ancient granites and plagiogranites. However, neither the relic granitic textures, nor intrusive contacts have been discovered in the region, and, consequently, the idea on the magmatic origin of parent rocks is only an assumption (though very probable). Accordingly, we attributed the described rocks to the metamorphic, and not plutonic unit. The Archean age of their protoliths is substantiated by the isotopic age value of 2700 Ma for zircons from migmatite-free gneisses of the Sortavala and Kokkaselkä domes (Tugarinov and Bibikova, 1980), which evidently corresponds to the upper age limit of their parent rocks.

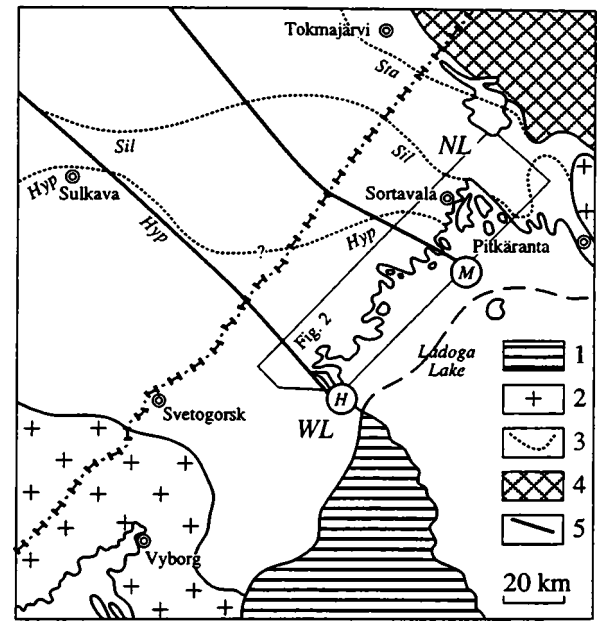

Fig. 1. Scheme of geographic distribution of metamorphic Ladoga Complex: (1) the Riphean–Paleozoic platform cover; (2) rapakivi plutons; (3) Early Proterozoic metamorphic belt with isograds of staurolite (*Sta*), sillimanite (*Sil*), and hypersthene (*Hyp*); (4) Archean complexes of the Karelian Massif; (5) Meiyeri (*M*) and Hiitola (*H*) faults; symbols *NL* and *WL* mark the North and West Ladoga zones respectively.

None of the West Ladoga formations can be reliably compared with the Archean granite-gneisses of the North Ladoga zone on the basis of their petrographic composition and textural features.

KARELIAN ROCKS

Lower Karelian (Jatulian and Ludicovian): Sortavala Group

In the North Ladoga zone, rocks of the Sortavala Group occur in narrow belts, rimming inliers of the Archean granite–gneiss basement. The group consists of two subgroups differing in composition. The lower subgroup of insignificant thickness (locally up to 200 m, but usually below this limit) is composed of metamorphosed clastic rocks, such as arkoses and quartzites, enclosing rare limestone beds. Some sections in the Ladoga region and eastern Finland show that the basal arkose horizon of the lower subgroup transgressively overlaps the granite-gneiss complex of the Archean basement (Sinitsa, 1984; Svetov and Sviridenko, 1992). The upper subgroup is much thicker (up to 2000 m), and indicative of the Sortavala Group. In the Ladoga region it consists of amphibolites interpreted as metamorphosed basic to acid volcanites (mostly tholeiitic basalts) and their tuffs, with a minor admixture of andesites and dacites. The most representative sections of the upper subgroup in edging structures of the Kir-

Subdivision scheme of the Lower Precambrian in the northwestern Ladoga Region

Stratigraphic subdivisions: general	Stratigraphic subdivisions: regional	Tectonic stages of Svecofennides evolution	Age, Ma: for stratigraphic boundaries	Age, Ma: for intrusions	Names of local subdivisions
Proterozoic, Karelian, Upper		Postorogenic		1760–1800	Ojajärvi complex of monzonite–granites (1)
		Late orogenic		1850–1860	Tervu complex of leucogranites (2)
		Synorogenic		1860–1870	Impiniemi subcomplex (3, 4); Lauvatsaari subcomplex (5) — Lauvatsaari–Impiniemi complex of metadiorite–tonalites; Synorogenic metagabbros and metanorites [Kalaamo, Welimjaki, and other massifs (7)]
				1870–1880	Kurkijoki complex of metanorite–enderbites (6)
	Kalevian	Initial	1900–1920		Ladoga Group (8) of metaturbidites (subdivided into formations); Lahdenpohja Group of gneiss–granulites (undivided). Metamorphic suites: biotite gneisses (11); garnet gneisses (12); diopside–biotite gneisses (13); diopside–garnet gneisses (14); hypersthene–garnet gneisses (15); hypersthene gneisses (16); cordierite gneisses (17); agmatites (18)
Proterozoic, Karelian, Lower	Ludicovian				
	Jatulian		1980 ± 27 2070 ± 10 2100–2120		Upper subgroup of metatholeiites (subdivided into formations); Subvolcanic complex; Amphibolite suite (10); Lower subgroup of metaclastic rocks — Sortavala Group (9)
	Sariolian and Sumian		2300		
Archean	Lopian		2500	2700	North Ladoga granite–gneiss association (19)

Ladoga metamorphic complex

Note: Numbers in parentheses correspond to symbol numbers in Fig. 2.

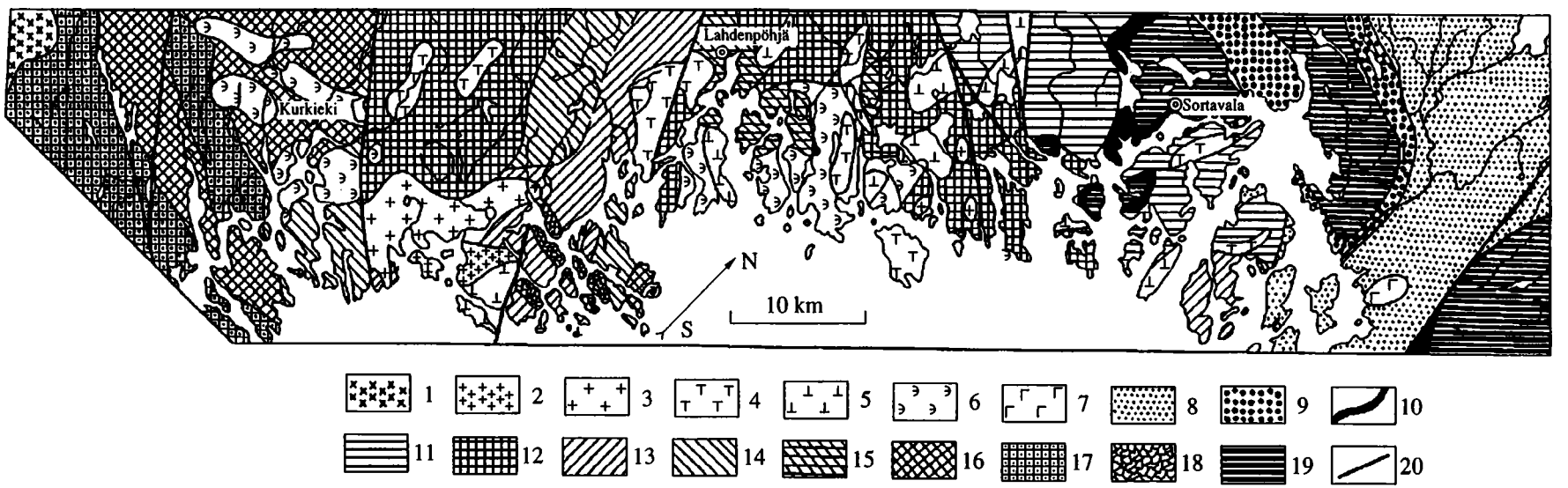

Fig. 2. Geological scheme of the northwestern Ladoga region (based on geological survey data of "Sevzapgeologiya" Co., and modified in accordance with our data): (1) postorogenic Ojajärvi complex of monzonite–granites; (2) late orogenic Tervu complex of metadiorite–granites; (3–7) synorogenic complexes: the Lauvatsaari–Impiniemi complex of dominant two-mica (3) and biotite to garnet–biotite (4) metatonalites of the early phase, and of late metadiorites (5); the Kurkijoki complex of metanorite-enderbites (6); and metamorphosed gabbros and diorites (7); (8) the Kalevian Ladoga Group of metaturbidites; (9–10) Jatulian + Ludicovian: the Sortavala Group of metatholeiites (9) and amphibolites (10); metamorphic suites: (11) biotite gneisses, (12) garnet gneisses, (13) diopside–biotite gneisses, (14) diopside–garnet gneisses, (15) hypersthene–garnet gneisses, (16) hypersthene gneisses, (17) cordierite gneisses, and (18) agmatites; (19) Archean: the North Ladoga granite–gneiss association; (20) faults.

javolahti dome have been studied in detail and subdivided into several formations by Svetov.

The Sortavala Group is associated with a subvolcanic complex of abundant dykes and sills, which occur not only in the volcanogenic sequence, but also in underlying rocks. Like the host rocks, many dykes are tholeiitic in composition, although more acid varieties have also been distinguished.

The problem of the original tectonic setting of the Sortavala tholeiites was studied by Svetov *et al.* (1990). Using geological and petrochemical criteria, these researchers demonstrated their affinity with plateau basalts. This conclusion conforms to our data and the results of Pekkarinen and Lukkarinen (1991), who showed that in chemical parameters metabasites of the Sortavala level correspond to the intraplate tholeiites.

Rocks of the Sortavala Group are so intensively metamorphosed and dislocated that relics of their premetamorphic textures and structures are preserved only in a few localities; it is usually difficult to infer the types of protoliths, their successions and original geologic relations. In the main part of the distribution area, the volcanogenic subgroup is transformed into monotonous schistose amphibolites. They are especially typical of the structural zones rimming the southernmost domes near Pitkjaranta and Sortavala, where amphibolites were extremely reduced in thickness (dozens of meters only) when rising granite-gneiss domes entirely distorted the original stratification of the rocks. Nevertheless, it is beyond doubt that metabasites located around granite-gneiss domes of the North Ladoga zone represent a single unit, the Sortavala Group. Despite the insignificant thickness, amphibolites and less altered schists, retaining features of their initial stratification, form the continuous rock slab, which overlies the granite-gneiss basement and underlies the metasedimentary Ladoga Group. These relations and overall morphology of the metabasite body attest that it is a volcanic blanket resting on the eroded surface of the basement.

In the West Ladoga zone, where outcrops of the Archean basement are lacking, reliable analogues of the Sortavala Group have not been recognized.

A particular problem lies in distinguishing amphibolite dykes among other rock types, because their original morphology is radically altered: as a rule, they are transformed into tectonically dismembered lenslike fragments of formerly intrusive bodies now oriented along the general schistosity. Nevertheless, in some localities, original features are preserved and show how dykes cut the schistosity and banding of the host rocks (Sinitsa, 1984; and others). In subvolcanic bodies, amphibolites are more coarse-grained than in surrounding metavolcanic rocks, especially when such bodies are relatively large. The tholeiitic composition of these dykes underscores their affinity to host metavolcanites, which is a typical of younger metabasites in the Ladoga region.

Thus, the Sortavala Group is divisible into the following units of different genesis: (a) the lower subgroup of metaclastic rocks; (b) the upper subgroup of metatholeiites constituting several formations; (c) the subvolcanic complex of metatholeiites; and (d) the amphibolite metamorphic suite, which should be admitted for geological bodies lacking original stratification or features of intrusive origin, and therefore incompatible with the above units.

For a long period, viewpoints on the age of the Sortavala and overlying Ladoga groups were controversial. Many famous researchers, such as Sederholm, who was the first to recognize the Ladoga complex in 1887, and Kratz (1963), who discriminated the Sortavala Group as an individual unit of this composite complex, argued for the pre-Jatulian age of both groups, and referred them to either the Archean or the basal Proterozoic. At present, however, there is a general agreement that the Ladoga Group is of the post-Jatulian age, whereas the Sortavala Group is either correlated with the Jatulian, or assigned to the Ludicovian (the subdivision right above the Jatulian interval) in accordance with the inference of Sokolov (*Materialy* ..., 1982; Svetov and Sviridenko, 1992).

The Ludicovian in the aforementioned Russian scale corresponds to the stratigraphic interval between the Jatulian and Kalevian. We assume the lower Ludicovian boundary to be 2070 ± 10 Ma old on the following basis: the age of definitely Jatulian ferruginous quartzites is 2080 ± 45 Ma (Vaasjoki *et al.*, 1980), whereas volcanic rocks of the upper Sortavala Subgroup, corresponding to the Ludicovian level in the opinion of Svetov and Sviridenko (1992), are 2062 ± 2 Ma old (Pekkarinen and Lukkarinen, 1991). However, the base of the upper Sortavala Subgroup may be of another age. The Jatulian–Ludicovian interval comprises several tholeiite horizons similar in composition and probably collateral to the Sortavala volcanics. It is unclear which of the horizons is compatible with the basal volcanic rocks of the Sortavala Group. The oldest manifestations of tholeiitic volcanism are likely distinguished in eastern Finland north of Tohmajärvi, where they are referred to the Koljola Formation with associated dykes. The dyke level is in the middle portion of the Jatulian sequence. The dykes cut the lower quartzite unit of the Jatulian, and are overlain by the upper quartzites. Zircon age determinations for dykes are within the range of 2100–2120 Ma (Pekkarinen and Lukkarinen, 1991). It is not inconceivable that initial episodes of the Sortavala volcanism are correlative with this inter-Jatulian level.

The time-range of the lower Sortavala Subgroup of metaclastic rocks is also unclear. Following the inferences of Makarova and Borisova, Svetov and Sviridenko (1992) suggest that the subgroup is a reduced analogue of the Jangozero and Tulomozero formations, i.e., corresponding in age to the entire Jatulian interval, or to its greater part.

The Sortavala Group is unconformably overlain by the Ladoga Group (*Materialy ...*, 1982). Svetov and Sviridenko (1992) argued for correlation between the upper part of the Sortavala Group and the Suisaari Formation, which is the terminal one in the Ludicovian succession of the Onega region. The Sm–Nd age of an ultramafic sill in the top portion of the Suisaari Formation is 1980 ± 27 Ma (Puchtel *et al.*, 1992), and this value can be considered as the minimum age limit of the Sortavala Group. If this is correct, then the accumulation of Sortavala intraplate tholeiites very likely ceased before a new epoch of tectonic instability suggested to be marked by the formation of the Outocumpy and Jormua ophiolites in eastern Finland, where they display an age interval of 1970–1960 Ma (Gaal and Gorbatschev, 1987; Pekkarinen and Lukkarinen, 1991).

KALEVIAN. LADOGA GROUP OF METATURBIDITES

In the North Ladoga zone, the Ladoga Group is widespread in the vast interdome areas. It is basically composed of metasandstones and metasiltstones, and has a generally terrigenous composition. These rocks display characteristic rhythms of graded bedding, evaluated to be a distinctive feature of turbidity sedimentation. The group is subdivided into a series of formations of unsound validity, and different scientific teams are adherents of their individual stratification schemes.

Rocks of the Ladoga Group are metamorphosed under the conditions of greenschist and epidote-amphibolite facies. They are represented by phyllites, mica and garnet-mica schists, as well as by staurolite-, andalusite-, and sillimanite-bearing schists and quartzites. In southern areas near the West Ladoga zone, the degree of metamorphism approaches conditions of the amphibolite facies, and here one can observe abundant migmatites and disappearance of the features of original rock stratification. The required criteria to refer these high-grade metamorphic rocks to the Ladoga Group are unavailable, and we consider them as constituting the metamorphic Lahdenpohja Group.

In Finland, metaturbidites similar to above-mentioned rocks are also widespread and referred to the Kalevian. One of their most complete sections is studied in the Tampere schist belt of southern Finland (Simonen, 1980; Kähkönen *et al.*, 1989), where turbidites mostly occur as basal beds of a volcanogenic-sedimentary succession. Volcanic rocks of this succession are represented by acid, intermediate, and basic lavas and tuffs of calc-alkaline affinity. They range in age from 1904 ± 4 to 1889 ± 5 Ma. The lower boundary age of the lowermost turbidite unit below the 1904-Ma-old volcanics is inferable from the age data for clastic zircons separated from sedimentary rocks and measured by SHRIMP. About two thirds of the 120 studied zircon grains showed the age range of 2.1–1.9 Ga, including particular values such as 1949 ± 42, 1938 ± 12, 1932 ± 30, 1915 ± 13, and 1907 ± 15 Ma (Huhma *et al.*, 1991). This indicates that turbidites are insignificantly older than volcanic rocks, and that, being age related, both rock types are constituents of a single island-arc complex.

The Tampere turbidites are classified as sediments of the frontal island-arc zone, while the Ladoga turbidites presumably represent deposits of the back-arc basin, as it follows from their proximity to the craton and from the lack of concurrent volcanogenic rocks. To judge by realistic models, the back-arc turbidites cannot be much older than the island-arc complex, whose age in the Svecofennian region is within the range of 1.91–1.87 Ga (Kähkönen *et al.*, 1989). We make an exception for the highly metamorphosed volcanic rocks of the Pielavesi locality (eastern Finland), which have not been dated by isotopic methods and underlie the Kalevian metaturbidites according to Ekdahl's observations (Ekdahl, 1993). This allows us to suggest that the locally exposed oldest granitoids of Svecofennides with an age range of 1.92–1.93 Ga (Vaasjoki and Sakko, 1988) are related to the last metavolcanics, and that precisely this rock association was a source for the clastic material constituting the Kalevian turbidites.

All these data suggest that in southern and eastern Finland, as well as in the Ladoga region, the basal boundary of Kalevian turbidites is not older than 1.91, or 1.92 Ga at best. In accepted stratigraphic schemes of the Precambrian, including those of Russia (Semikhatov *et al.*, 1991), the basal Kalevian boundary is placed at an older level of about 2 Ga, while the well-traceable 1.9-Ga-old boundary separates the Lower and Upper Karelian and correlates with the top of the Kalevian. The refined age of the Svecofennian events changes this scheme: the main event, which is well marked and traceable in the Baltic Shield at the level of 1.9 Ga and should be acknowledged as boundary separating the Lower and Upper Karelian, was an incipient accumulation of island-arc volcanic and turbidite complexes, and not the outbreak of plutonism and metamorphism that terminated this event 1.88–1.86 Ga ago. This is a good reason to consider the Kalevian as the lowermost subdivision of the Upper Karelian comprising age-related volcanic and turbidite complexes of the early evolutionary stage of Svecofennides, and we are not the first advocates of this opinion (see Bibikova *et al.*, 1989). In this case, a hiatus ranging 50–60 Ma and preceding the turbidite accumulation in the North Ladoga region should be considered as corresponding to the Late Ludicovian time.

UNDIVIDED LUDICOVIAN AND KALEVIAN. LAHDENPOHJA GROUP OF GNEISS-GRANULITES

The Lahdenpohja Group includes intensively metamorphosed rocks of the North and West Ladoga zones, which are especially abundant in the latter. Preserved relics of layered rock textures and abundance of high-Al rocks, associated with quartzites and the carbonates,

suggest the sedimentary origin of the group. Locally it comprises rocks of presumably volcanic, but not very clear genesis. Some small bodies probably are composed of metaintrusive rocks. Despite the abundance of originally stratified rocks, it is impossible to understand the structure of the original sequence with presently available data, because essential lithologic features are preserved in fragmented and separated localities, thus complicating the identification of beds and their successions. We failed to distinguish the sedimentary origin of boundaries for most of the geological bodies revealed by mapping.

Nevertheless, we differentiated the Lahdenpohja Group, heterogeneous in composition and structure to constituent metamorphic suites, or *petroformations* (Shuldiner, 1992). These suites represent paragenetic associations of metamorphic rocks, which can be first of all distinguished by means of such a criterion as the co-occurrence of the rocks. During our geological survey, we identified several suites with different rock associations. The applied suite names are intended to denote the most characteristic rock types of the respective associations; a combination of two rock names means that alternation of these rocks is the most notable feature of the suite. Below, there is a brief description of the distinguished suites in a succession, illustrating the growth of their distribution areas from northern to southern regions (Fig. 2).

The Biotite gneiss suite mostly consists of biotite plagiogneisses and schists displaying a relatively low degree of migmatization. Associated rocks are represented by garnet–biotite gneisses, and by gneisses and schists with variable assemblages of paragenetic minerals depending on the metamorphic grade (sillimanite-, andalusite-, and cordierite-bearing gneisses, two-mica and chlorite–mica gneisses and schists). Outcrops of the biotite gneiss suite are concentrated in the North Ladoga zone.

The Quartzite suite of limited distribution is exposed in the North Ladoga zone as thin (dozens of meters) and scattered lentils of either monomineral, or biotite, garnet, and graphite quartzites associated with biotite and biotite–garnet plagiogneisses.

The Amphibole–garnet gneiss suite includes an association of hornblende gneisses and amphibolites alternating with biotite and biotite–garnet plagiogneisses. The rocks are exposed in a series of small outcrops along the boundary between the North and West Ladoga zones.

The Garnet gneiss suite is very typical of both metamorphic zones of the Ladoga region. It mostly consists of garnet–biotite plagiogneisses and migmatites associated with scarce representatives of other rock types. The suite is monotonous in composition, but very diverse in structure because of irregular migmatization (lit-par-lit, reticulate, porphyroblastic, and other migmatites).

The Diopside–biotite and diopside–garnet gneiss suites differ from the biotite and garnet gneiss rock assemblages because of the associated calcareous rocks, such as diopside schists and gneisses, or calcite- and scapolite-bearing and amphibole schists, and rare marbles. The calcareous schists and gneisses occur as thin interlayers in biotite and garnet gneisses, or often as rounded zonal inclusions resembling concretions typical of schists of the Ladoga group. This misleading similarity was sometimes used to make the assumption that the suites are correlative with particular subdivisions of the Ladoga Group. However, our observations showed the inclusions to be mostly of tectonic origin. They consist of fine-grained nonmigmatized rocks, which are more ductile than the host gneisses and easily yielding boudins under the extension. Their zoning is a result of a metamorphic interaction between rocks of different composition, which is also typical of interlayers.

The Hypersthene–garnet gneiss suite of limited distribution occurs in the West Ladoga zone near the Lahdenpohja. Its characteristic feature is a regular flyschlike interlayering of biotite, hypersthene–biotite, and hypersthene–garnet–biotite plagiogneisses that slightly resembles rock alternations in the Ladoga Group.

The Hypersthene gneiss suite is widespread in the southern part of the West Ladoga zone. The most abundant hypersthene–garnet rocks of the suite are locally alternated with garnet- and hypersthene–biotite gneisses; however, this interlayering is less regular than in the foregoing suite.

The Cordierite gneiss suite is also very typical of the southern West Ladoga zone. It includes high-Al plagioclase and two-feldspar gneisses and schists showing variable percentages of cordierite, garnet, sillimanite, and biotite. Some rock varieties correspond to garnet–biotite gneisses. The intense migmatization is typical of all rock types.

Agmatites represent a particular suite type identified on the basis of the structures, and not the compositions of the rock associations. This suite comprises large-cell reticulate migmatites (polymigmatites in Simonen's terminology) with breccia-, or even conglomerate-type structural patterns. Leucosomes of these rocks dominate over the lit-par-lit migmatized matrix, whose fragments are chaotically oriented. Agmatites occur in vaguely bounded zones, which are up to 2 kilometers wide and several kilometers long, and crosscut fields of other suites. The petrographic features of agmatites depend on the matrix composition. Their origin seems to be related to more intense migmatization in extension zones.

As one can see from the above description, the petrographic composition and general outlook of metamorphic suites depends on several factors, such as original rock lithology, thermodynamic conditions of metamorphism, and tectonic environments. The distinguished suites reflect the integrated influence of

all factors, and the zonal distribution of the suites affords an opportunity to judge how these factors have varied throughout the studied area. For instance, embodiments of calcareous diopside-bearing rocks inside metapelites illustrate the specific lithology of protoliths, while the appearance of hypersthene-bearing suites indicate the higher grade of metamorphism. However, in certain cases, factors responsible for petrographic differences, e.g., between the biotite and garnet gneiss suites, are not so easily understandable, and can be interpreted only by means of a special analysis.

The substratum age of the Lahdenpohja Group can be inferred only within a wide range. In the North Ladoga zone, we observed clear indications of a gradual transition between the low-grade metamorphic rocks of the Ladoga Group and high-grade metamorphic associations of the Lahdenpohja Group which suggests their synchronism. This correlation is also based on the similar occurrence of both units above rocks of the Sortavala Group surrounding the granite–gneiss domes. In the West Ladoga zone, these signs of correlation are erased. However, we did not note any sharp difference in the Lahdenpohja Group composition at the boundary between the two zones, and consequently, at least a part, or even the greatest part of high-grade metamorphic rocks of the West Ladoga zone can be correlated with the Ladoga Group. Rock relics with characteristic graded bedding, which are preserved among the metapelites of this zone, favor such a conclusion. On the other hand, the Lahdenpohja Group here is more diverse in composition than the Ladoga Group, which may indicate the wider time range of the former. It is clear, however, that this group does not include the post-Kalevian rocks, because it is cut by the early orogenic intrusions with the age of 1.88–1.87 Ga (see below). Its lower age limit in the West Ladoga zone is still unclear, because granite–gneiss domes, mantled by easily recognizable Sortavala amphibolites, are absent here. Consequently, we cannot identify the Kalevian base in this area, and its metamorphic succession may include older formations. Provided that the pre-Ludicovian rocks in Svecofennides have never been detected by isotopic methods, we may refer the Lahdenpohja Group to the undivided Ludicovian–Kalevian interval.

LATE KARELIAN INTRUSIONS

Plutonic rocks that are related to the synorogenic (1880–1860 Ma), late orogenic (1860–1850 Ma), and terminal postorogenic (1800–1760 Ma) stages of Svecofennides evolution are of the Late Karelian age. Their subdivision was first substantiated by Saranchina (1972), and further detailed in subsequent works (*Migmatizatsiya ...*, 1985; Kotov and Samorukova, 1990; Shinkarev *et al.*, 1992; Shuldiner *et al.*, 1992; and others).

Metaintrusive Complexes of Synorogenic Stage

Metagabbros and metanorites. Synorogenic metaintrusive rocks of basic composition represent the early phases of the following Kurkijoki and Lauvatsaari–Impiniemi complexes, as they occur as xenoliths in their late phases. In addition, metagabbroids, metanorites, and sometimes ultramafic rocks occur in the form of individual bodies, usually of a small size (from less than meter to several meters across), although some of them are significantly larger, e.g., the Kalaamo massif of norite–gabbronorites and the Velimäki massif of ultramafite–gabbro–diorite composition (Saranchina, 1972; Svetov *et al.*, 1990). As a rule, it is impossible to identify their affinity with any one of above complexes; however they proved to be older than tonalites of the Impiniemi subcomplex.

The Kurkijoki complex of metanorite–enderbites. Rocks of this complex are concentrated in the Lahdenpohja and Kurkijoki localities (Saranchina, 1972; Kotov and Samorukova, 1990). They are exposed over separate areas, some of which are up to 40–50 km^2, or as compact groups of small outcrops, which may represent fragments of large bodies separated by narrow roof sags or tectonic slices of enclosing rocks. Small enderbite bodies covering areas of up to 1–2 km^2 are scattered all over the West Ladoga zone, but are absent in the northern territory. The rock composition varies from norites to hypersthene plagiogranites; therewith, basic rocks are more abundant in the Lahdenpohja locality, whereas acid varieties are typical of the Kurkijoki exposures. The U–Pb zircon ages for hypersthene granitoids of the Ladoga (Kurkijoki complex) and nearby regions of Finland (Kiuruvesi–Haukivesi complex) are within the range of 1880–1870 Ma (Tugarinov and Bibikova, 1980; Korsman *et al.*, 1988); in particular, the Kurkijoki massif is 1871 ± 6 Ma old (Kotov *et al.*, 1992).

The garnet leucogranites, which sporadically occur among the Lahdenpohja enderbites, e.g., in the Vavasaari locality, seem to also represent the early metaintrusive rocks (Svetov and Sviridenko, 1992). These rocks, unusual in the Ladoga region, have a high content of K-feldspar, but are extremely depleted in biotite. This probably resulted from the parent magma composition, which was highly undersaturated with water, and had a high FeO/FeO + MgO ratio.

The Lauvatsaari–Impiniemi complex of metadiorite–tonalites. This complex is divisible into two phases (besides the early gabbroids), which are described as individual manifestations of magmatic activity because of their considerable difference in composition (*Migmatizatsiya ...*, 1985; Shinkarev *et al.*, 1992). Based upon these data, we consider them in the rank of subcomplexes using a combination of traditional names to designate the complex as a whole. Both subcomplexes are widespread throughout the Ladoga region, being especially abundant in the West Ladoga zone, where they form outcrops up to 20–30 km^2 in area.

The Lauvatsaari subcomplex consists mostly of biotite–hornblende metadiorites and quartz diorites that intruded into rocks of the Lahdenpohja Group, including enderbitiods. The main rocks of the Impiniemi subcomplex are biotite, two-mica, and local garnet–biotite metatonalites. Rocks of tonalitic composition sometimes grade into quartz diorites, but locally intrude them. The intrusive contacts between tonalites and all of the older rocks of the Ladoga region are unquestionable. The U–Pb zircon age of the Lauvatsaari massif is 1864 ± 13 Ma (Kotov *et al.*, 1992), and taking into account the polyphase structure of the complex as a whole, we believe that it was formed about 1870–1860 Ma ago.

Available data on major and trace element geochemistry suggest that the Kurkijoki and Lauvatsaari–Impiniemi complexes are the crystallization products of low-K high-Ca magmas of the I-type, characteristic of magmatic-arc zones. Rock series of both complexes display the calc-alkaline trends of magmatic differentiation. Distribution patterns of major and trace elements (including the rare earth elements) show that parent enderbite and gabbro–tonalite magmas were generated in a mantle reservoir with the additional influence of crustal components.

The Metaintrusive Complexes of Late Orogenic Stage

The Tervu complex of metadiorite–granites includes several generations of dioritic and granitic massif and small massifs. The largest granitic massif (4 × 6 km) near Tervu represents a relatively thin (hundreds of meters) subhorizontal body underlain by a thick zone of eruptive breccias and dykes (Shuldiner *et al.*, 1995).

The early dykes of mostly dioritic composition crosscut the metamorphic, often migmatized rock in different directions relative to the rock banding. The dykes themselves are metamorphosed into amphibolites and biotite–hornblende–plagioclase, or biotite–plagioclase schists, many of which are also affected by migmatization. The schistosity of metamorphic dykes is parallel to their contacts crossing the host rock foliation. The dykes experienced dislocation and folding; some folds are compressed up to isoclinal forms. The fact that the dykes crossing metamorphic rocks are also metamorphosed suggests two stages of rock transformations. Age relations between these stages are illustrated by metadiorite dykes crossing metatonalites of the Impiniemi subcomplex.

The latest phases of the Tervu complex are represented by two-feldspar granites and trondhjemites displaying a high degree of muscovitization. These granitoids are intensively deformed, and show blastomylonitic structures everywhere. The rocks have clear eruptive contacts with all of the older rocks, including metadiorite dykes. Shinkarev *et al.* (1992) classified these rocks as moderately alkaline granitoids of the crustal S-type characteristic of collision zones. The U–Pb zircon age of the Tervu granites is 1856 ± 6 Ma (Kotov *et al.*, 1992). In southern Finland, similar late orogenic granitoids yielded zircons with the age range of 1.83–1.81 Ga (Vaasjoki and Sakko, 1988).

POSTOROGENIC INTRUSIONS

Ojajärvi Complex of Monzonite-Granites

The postorogenic formations of the Early Proterozoic comprise a series of massifs and dykes unaffected by plastic deformations and metamorphism. They are most typical of the West Ladoga zone, where their abundance is growing in a southward direction. Thus far, the systematics and subdivision scheme of postorogenic intrusions are imperfect.

The Ojajärvi massif of subalkaline granitoids is a typical representative of such intrusions. This massif, 15 × 6 km in dimension, is oriented parallel to the meridian direction and discordant relative to metamorphic rocks of the Lahdenpohja Group. The massif bears xenoliths of host rocks, and shows clear intrusive relations with them along outer boundaries. Marginal zones of the massif are contaminated by the host rock material, while its inner uncontaminated part consists of purely magmatic rocks, such as monzonitic gabbrodiorites, monzonites, quartz syenites and granites. The intrusion is of a multistage origin (Ladner, 1983), and according to preliminary data, was formed 1.80–1.76 Ga ago (Shinkarev *et al.*, 1992).

Afterwards, intense magmatism of a different type was activated during the very late Karelian–very early Riphean epoch to give rise to the large rapakivi plutons, including the Wiborg massif with measured age values of 1646–1630 Ma (Vaasjoki and Rämö, 1989).

Metamorphic Zones

We consider metamorphic zones to be specific geological subdivisions distinguishable by means of such criteria as the thermodynamic conditions of metamorphism, which affected the rock groups under examination, whereas their chronological and genetic parameters are ignored. Mapping the metamorphic zones, we may understand the T–P structure of metamorphic complexes. Boundaries of metamorphic zones can be depicted on the basis of various indications; the most simple sign is the appearance of index-minerals, or their typomorphic paragenetic assemblages in the rocks which are detectable during field observations. It is a common practice to distinguish the garnet–staurolite–sillimanite–muscovite, sillimanite–K-feldspar, and hypersthene zones. In the Ladoga region, the metamorphic zoning was studied by Velikoslavinskii (1972) and Nagaitsev (1974), while in nearby areas of Finland, it was investigated by Korsman *et al.* (1984, 1988). Despite the somewhat schematic and fragmentary state

of the available data, they still reveal the sublatitudinal to northwest striking patterns of general metamorphic zoning, which shows an increase of metamorphic grades in going from the northern to southern zones, the latter being more distant relative to the Karelian Massif (Fig. 1).

The least clear is the configuration of the hypersthene zone, because retrograde alterations resulted in the replacement of this very unstable mineral phase. In the West Ladoga zone, the relic assemblages are known only in the Lahdenpohja and Kurkijoki localities, and previous schemes show two separate fields of the hypersthene zone. More detailed investigations revealed hypersthene relics within a considerably larger territory extending from the Suri-Heposaari Island in the northeast (Sortavala region) to the boundary of the Wiborg rapakivi massif in the southwest; over this entire area, we also encountered the enderbite bodies. The northeastern boundary of the granulite distribution is detected on the basis of such indications as the replacement of relic hypersthene (or secondary fibrous cummingtonite) by well-edged crystals of metamorphic cummingtonite. The hypersthene isograde is also well traced west of the Russia–Finland border, along the northern granulite outcrops near Sulkava (Korsman *et al.*, 1984, 1988). However, it is unclear as to whether or not the Sulkava and Ladoga granulite fields join one another.

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Some Aspects of Magnetostratigraphic Correlation

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Abstract—The traditional visual comparison of paleomagnetic columns with the consecutive counting of magnetozones of different polarities is liable to many errors. In this approach the following methodical miscalculations are particularly apparent: the loss of an interval, an arbitrary interpolation of paleomagnetic boundaries, comparison of magnetozones of different ranks, and erroneous identification of magnetozones. Paleomagnetic stratigraphic correlation is practically impossible without exact correlation between paleontological, lithological and facies data. Appropriate examples are given to show the procedure of magnetostratigraphic correlation of marine and continental deposits in widely spaced biogeographic provinces.

Key words: *magnetostratigraphy, magnetozone, polarity of remnant magnetization, correlation, paleomagnetizm, scale intervals, inversion.*

INTRODUCTION

The correlative capacity of the magnetostratigraphic method was highly appreciated by many specialists after the first publication of the results of a paleomagnetic study of the Permian–Triassic and Pliocene parts of the scale. But as the age span of paleomagnetic determinations extended and the data interpretations became more exact, the magnetostratigraphic speculations began to lose their seeming uniqueness and often lead to uncorrelated and even contradicting conclusions.

In this connection, particularly demonstrative is the long standing discussion about the "short" and "long" Pontian Stage and, consequently, about its position in the general magnetochronological scale. By using, on the whole, similar paleomagnetic characteristics of a series of the Black Sea sections, different authors correlate the Pontian with the lower portion of the Gilbert zone, magnetic epoch 6, and even with the base of the Gilbert zone plus polarity epochs 5 and 6 (Semenenko and Pevzner, 1979; Chepalyge *et al.*, 1985; Zubakov, 1990).

The stratigraphic position of the boundary between the Kiama and Illawarra paleomagnetic hyperzones was represented differently. In European Russia, this boundary was placed in the middle of the Tatarian Stage of the upper Permian (Khramov, 1963; Molostovskii, 1983), whereas for the southwestern regions of the United States it is shown in the lower Triassic (McMachon and Strangway, 1968).

These kinds of discrepancies are quite common in magnetostratigraphic practice, and I need not enumerate them. The present situation is far from accidental, being a corollary of both the specific peculiarities of the paleomagnetic method itself and of the numerous methodical errors in the final interpretation of data.

The magnetozones determined according to only one criterion, i.e., the polarity of natural remnant magnetization (NRM), practically lack distinct individual features, thus greatly impeding their identification in individual sections and the paleomagnetic correlation itself.

The analysis of numerous publications reveals that the correlation procedure is usually reduced to a visual comparison of columns of magnetic polarity in definite stratigraphic intervals and to counting the zones of normal and reversed polarity in terms of their temporal succession. In principle, this kind of correlation is permissible; its correctness, however, can be vouched for only when certain conditions well known to stratigraphers but often ignored by geophysicists are rigorously observed.

A list of the most typical magnetostratigraphic errors was discussed in an earlier publication (Molostovskii and Khramov, 1984). The greatest distortion of paleomagnetic correlations occurs as a result of the following methodical miscalculations.

1. The loss of a stratigraphic interval can misrepresent the whole structure of the paleomagnetic column by merging magnetozones of different ages.

2. An error may occur if the principles of outright rejection and conditional assumption, which form the basis of any magnetostratigraphic construction, are ignored. This implies that the rocks with different NRM polarity are always of different ages, while the same NRM polarity does not prove the layers to be correlative.

3. The violation of the rule of correlating magnetozones of the same rank in accordance with their stratigraphic volumes can lead to mistakes, such as the establishment of a correlative framework having an insufficient number of reference magnetozones with reliable age allocations.

4. Errors can be caused by arbitrary interpretations of paleomagnetic boundaries.

The enumerated methodical miscalculations lead to the erroneous identification of magnetozones, which

has become the Achilles' heel in the paleomagnetic approach to stratigraphic correlation.

As the vast experience of magnetostratigraphic research testifies, autonomous paleomagnetic correlations are possible only under the exclusively favorable conditions of geologic and paleomagnetic factors such as complete sections, reliable paleontological or radiochronological datings, the paleomagnetic stability of rocks, and the distinct paleomagnetic structure of columns. Since, in practice, such combinations are extremely rare, magnetostratigraphic correlations usually become complicated, long processes of careful coordinaton of paleomagnetic, paleontological, lithological and facies data.

The results of a series of magnetostratigraphic interpretations of the upper Permian and lower Triassic formations of Eurasia illustrate this process.

MAGNETOSTRATIGRAPHIC CORRELATION OF THE UPPER PERMIAN FORMATIONS IN EUROPEAN RUSSIA, THE PAMIRS, AND TRANSCAUCASUS

The first data on the character of the magnetic zonality of the upper Permian were obtained on the reference sections of red rock formations in European Russia and in the southwestern United States. The regional magnetostratigraphic scale, developed by Khramov (1963), was confirmed by subsequent research; it is now accepted as the world paleomagnetic standard. Its reliability is provided by the completeness and ample paleontological description of numerous reference sections and the sufficiently precise allocations of magnetozones to the subdivisions of the general and regional scales.

The large R–Kiama and NR–Illawarra hyperzones compose the basis of the reference scale; their boundary coincides with the base of the upper Tatarian Substage. On the whole, the section of the Russian upper Permian contains five paleomagnetic orthozones; i.e., R_1P, N_1P, R_2P, N_2P, and R_3P. The lower zone of reversed polarity (R_1P) at the top of the Kiama hyperzone unites the Ufimian, Kazanian and lower Tatarian stages. The next four zones are restricted by the upper Tatarian Substage and, in the general paleomagnetic scale, correspond to the base of the Illawarra hyperzone (Khramov, 1963; Molostovskii, 1983).

The reference zones are complicated by the presence of several narrow subzones with the opposite sign; their location can be determined only in some of the sections because of the local erosions and paleomagnetic instability of some of the rocks. There are at least two subzones with reversed polarity in the N_1P zone, and one in the N_2P zone; there are also indications of the presence of the N-subzone in R_3P orthozone.

The Permian paleomagnetic scale of the Russian plate is well ranked and cognizantly generalized up to the level of orthozones; these characteristics make its application reliable over vast territories.

In the 1970s and the 1980s, Permian deposits of the Tethys were paleomagnetically studied in the reference sections of the Pamirs and Transcaucasus regions as the major objects of research. Later, the paleomagnetic column of the Tatarian Stage in European Russia was compared with the magnetic polarity in the Permian deposits of the Tethys. In these correlations, the boundary between the R–Kiama and NR–Illawarra zones served as a chronological marker; according to current data, in the sections of the central Tethys this boundary is confined to the base of the Midian Stage (Kotlyar *et al.*, 1984; Davydov *et al.*, 1982).

Since the Permian portion of the Illawarra sign-variable zone in continental formations refers to the upper Tatarian Substage, the Midian and Djulfian stages are now considered as an analogue to that substage in the Tethyan marine sections. Based on the different number of magnetozones in the Russian plate and Transcaucasus, Kotlyar, Sokarev, and a number of other researchers have concluded that the upper Permian section lacks the stratotype analogues of the Dorashamian Stage in its complete volume (Kotlyar *et al.*, 1984; Zakharov and Sokarev, 1991; Davydov *et al.*, 1982).

Goman'kov (1991) holds a similar viewpoint that the position of analogues of the Tatarian Stage in Western Europe and the Tethys is not the highest in the scale, and that these analogues are separated from the Triassic by a time interval of about two stages.

These views on the relations between the upper Permian continental and marine formations have become popular among Russian researchers as of late. Since these opinions are based to a large extent on paleomagnetic determinations, the latter acquire particular importance. The following circumstances have become apparent in the course of an analysis of the paleomagnetism of Permian deposits in the Pamirs and Transcaucasus.

It is absolutely obvious that the actual substantiation of the summarized Permian paleomagnetic column of the Tethys is inadequate, being a collection of separate zones from individual sections, many of which are identified by only one or two samples. Moreover, owing to the intense remagnetization of rocks in some of the sections, no more than three rock samples could be used for allocations. Consequently, the magnetozones of the Tethyan scale are not ranked, their boundaries are arbitrarily interpreted and, further, they represent a combination of individual samples distributed at random over the time scale.

Naturally, as it stands, the calculation of magnetozones in the Tethyan sections and their comparison with the magnetozones of the Tatarian Stage of the Russian plate lack any real significance. Moreover, the magnetostratigraphic scale of the upper Tatarian Substage has several subordinate subzones in addition to the four reference zones. Therefore, the total number of

inversions recorded in the Tatarian Stage can be easily correlated with that in the upper Permian of the Tethys area. This circumstance makes the basic conclusions about the comparative completeness of the Permian sections in these provinces questionable, even if we overlook the problem of the quality of paleomagnetic materials on the southeastern Pamirs and Transcaucasus.

A HIATUS BETWEEN THE TRIASSIC AND THE PERMIAN ON THE TERRITORY OF THE RUSSIAN PLATE

The upper Permian and Triassic formations of Laurasia are divided by a large depositional hiatus, which is particularly evident in the areas of continental sediments. There are different points of view concerning the age interval of the hiatus. Some authors place it at the end of the Permian, while others refer its larger part to the onset of the Triassic.

The paleomagnetic studies of the Permian and Triassic boundary units carried out in the eastern part of the Russian plate and in the Urals foredeep show that the problem is wholly impracticable, because the available data indicate that the duration of the hiatus widely varies, and in concrete geologic environments it is adjacent to different parts in both systems.

In the northern regions of European Russia, the lower Triassic horizons, equivalent to the N_1T magnetozone, are missing. In the section of the drainage basin of the Malaya and Bol'shaya Severnaya Dvina, the upper part of the Vyatka Horizon (R_3P) is contiguous to the Triassic R_1T zone, the lower strata of which are also reduced (Fig. 1).

A similar range of the hiatus was supposed in the Northern Urals (Molostovskii, 1983), but recent finds of procolophonids *Tichvinskia*(?) and theconodonts of the Erytrosuchidae family in the lower beds of the Byzovaya Formation (Novikov *et al.*, 1992) refer these deposits to the R_2T zone of the upper Triassic horizons and not to the R_1T zone. These data provide the grounds to presume that zones N_1T, R_1T, and N_2T, constituting a large part of the lower Triassic, are missing from the section of the Pechora River basin (Fig. 1, *2*).

In the basin of the middle course of the River Mezen', the hiatus is mostly confined to the Tatarian Stage of the Permian, where only two zones, R_1P and N_1P, are apparent. The Triassic portion of the scale for this area lacks the N_1T and, partly, R_1T zones (Fig. 1, *3*).

The southern part of the Urals foredeep shows a relatively complete lower Triassic sequence and the hiatus coincides with the Vyatka Horizon of the Permian, where zone R_3P is reduced (Fig. 1, *4*).

The minimal amplitudes of the Permian–Triassic hiatus are recorded on the southeastern margin of the Russian plate (the Volga region, the Southern Urals), where the lower Triassic magnetozone in the section joins the R_3P zone at the top of the Vyatka Horizon (Fig. 1, *5*).

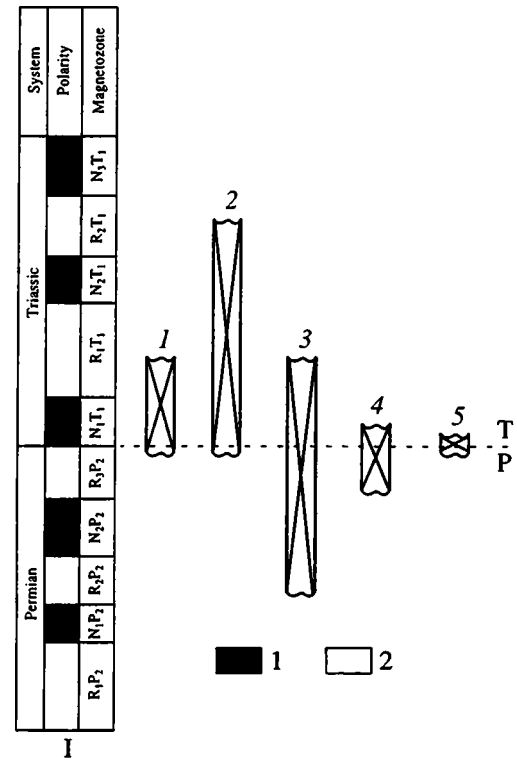

Fig. 1. The duration of the hiatus between the Triassic and Permian as indicated by paleomagnetic data (I-composite paleomagnetic section of the Upper Permian and Lower Triassic of the Russian plate). 1—Normal polarity; 2—reversed polarity.
(*1*) Bol'shaya Severnaya Dvina basin; (*2*) Pechora basin; (*3*) Mezen' basin; (*4*) Orenburg–Urals region (the foredeep zone); (*5*) Orenburg–Urals region (the platform part).

To conclude, it is obvious that the solution of the problem of the stratigraphic hiatus at the Paleozoic–Mesozoic boundary is far from simple, even within one region. This circumstance should not be disregarded in distant correlations and, in particular, when discussing the problem of comparative completeness of the Permian sections of the Tethys and of the Russian plate.

PALEOMAGNETIC CORRELATION OF THE LOWER TRIASSIC MARINE AND CONTINENTAL FORMATIONS

The problem of the correlation of stratigraphic scales of different biogeographic provinces is particularly important for the lower Triassic subdivision, in which marine and continental formations occupy commensurable areas. The stratigraphy of the marine lower Triassic is traditionally based on ammonites, but its stage subdivision has yet to be universally approved. Three individual zonal schemes are now compiled for the Tethyan and Boreal regions. The stratigraphy of the lower Triassic continental formations is based on the fauna of land vertebrates, and over the last few years the palynological data have acquired an increasing importance in its substantiation.

Not long ago, the potentialities of magnetostratigraphy in correlating the tetrapod and ammonite scales were limited by insufficient paleomagnetic data on marine standards. The research carried out in recent years in Arctic Canada and southern China has filled this gap and created a reliable basis for transcontinental correlations (Fig. 2).

The paleomagnetism of red non-marine formations of the lower Triassic was studied in European Russia, Germany, and in southwestern regions of the United States. The most detailed research was carried out in the Russian Triassic deposits by studying mutually correlated sections with sound paleontological characteristics. In total, the paleomagnetic column of this vast region has five orthozones, i.e., N_1T, R_1T, N_2T, R_2T, and N_3T, with several narrow subzones (Molostovskii, 1983). The largest subzone is documented, probably, within the N_2T zone of an interval corresponding to the Ust'-Myla Horizon (Novikov, 1991).

The correlation of magnetozones with the horizons of the regional scale and different groups of fossils is shown in the table.

The magnetostratigraphic scale of the lower Triassic for European Russia has no analogues with regard to the completeness of paleontological information; this makes it ideal as a reference scale for the continental lower Triassic. Separate elements of the scale are readily recognizable in the composite sections of western Europe and the United States.

In the German basin, all facies varieties of the mottled sandstone display four magnetozones (Nairn, 1960; Dachrot, 1976). The lower sandstone is bizonal, but some regions lack the first N-zone in the section, and the Zechstein is overlain by the reversely magnetized Broeckel Beds.

The higher horizons of the mottled sandstone comprise two magnetozones of normal and reversed magnetization. The lower N-zone includes the upper strata of the lower and middle sandstone units; the overlying R-zone belongs to the Rhaetian. The stratigraphic succession of magnetozones of the German Triassic section suggests them to be correlative with zones N_1T_1, R_1T_1, N_2T_1, and R_2T of the Russian sections.

These successions imply correlation between the lower sandstone and the Vokhma, Rybinsk and Sludka horizons, between the middle sandstone and the Fedorovka Horizon, and between the Rhaetian and the upper portion of the Fedorovka and the lower part of the Gam horizons of the Russian plate. Comparative analysis of palynocomplexes and conchostracans from the lower sandstone bed and the Vokhma Horizon and the protosuchians from the middle sandstone bed confirms the paleomagnetic correlations.

Individual correlations of paleomagnetic determinations in the Chugwater and Moenkopi red formations of

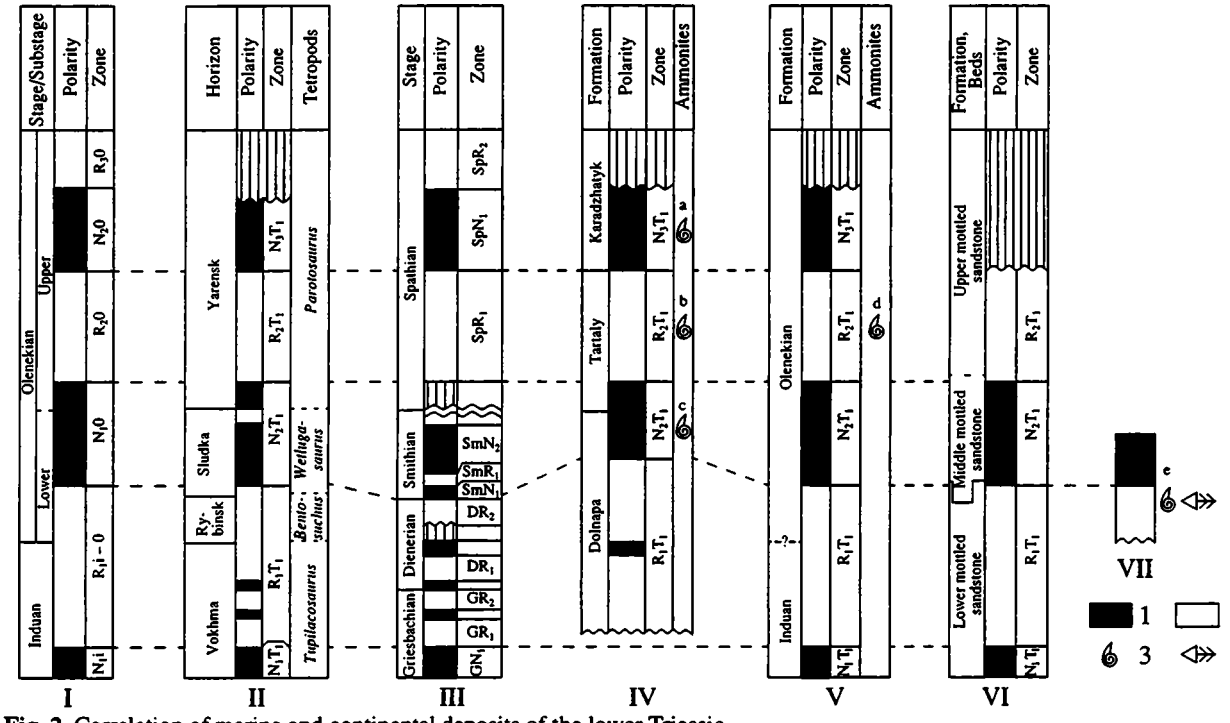

Fig. 2. Correlation of marine and continental deposits of the lower Triassic.

1—normal polarity; 2—reversed polarity; 3—ammonites; 4—tetrapods.

(I) General magnetostratigraphic scale. Regional scales: (II) European part of Russia (Molostovskii, 1983); (III) Arctic Canada (Ogg and Steiner, 1991); (IV) Mangyshlak (after Slautsitais personal communication); (V) Taimyr (Gusev, 1975); (VI) Germany (Nairn, 1960; Dachrot, 1976); (VII) Caspian region, Bol'shoe Bogdo Mt. (Molostovskii, 1983). (a) *Tirolites, Dorichranites;* (b) *Columbites;* (c) *Stacheites;* (d) *Olenekites, Keyserlingites;* (e) Burial of coupled *Protosaurus* and *Doricranites bogdoanus.*

Correlation of magnetozones with the lower Triassic regional subdivisions of the Russian plate

Horizon	Paleomagnetic column	Tetrapods	Dipnoans	Charophytes
Gam	N_3T_1	*Parotosaurus*	*Ceratodus*	*Auerbachichara bascuntscacensis*
	R_2T_1			
Fedorovka				
Ust'–Myla	N_2T_1	*Wetlugasaurus*	*Gnathorhiza + Ceratodus*	*Porochara triassica*
Sludka				
Rybinsk	R_1T_1	*Bentosuchus*	*Gnathorhiza*	*Altohara continua*
Vokhma	N_1T_1	*Tupilacosaurus*		*Wladimiriella wetlugensis*

the southwestern United States are the result of insufficient paleontological data for their sections. The North American sections are obviously lacking the first Triassic N_1T zone, and the R_1T zone here is coupled with the upper part of the Kiama R-zone. Apparently, the loss of a stratigraphic interval misplaced the position of the Kiama–Illawarra hyperzones boundary inside the lower Triassic (McMachon and Strangway, 1968).

The Dean wood and Goose Egg formations, displaying stable reversed polarity, correspond to the R_1T zone and, probably, to the upper half of the Vokhma Horizon. These formations contain beds with *Claraia* and ammonites of the *Otoceras* Zone which favors the above correlation. The reversely magnetized Virginian sandstones with *Tirolites* and lower beds of the middle red member in the Colorado sequence can be correlated with the R_2T_1 zone of the Russian scale (Lozovsky and Molostovsky, 1993).

The sections from the Mangyshlak Mountains, Taimyr (the Tsvetkov Cape), and Arctic Canada are the most convenient for correlation of the lower Triassic marine and continental formations. These sections have a distinct structure of magnetic reversals that is favorable for reliable correlations between the ammonite and paleomagnetic zones (*Paleomagnetizm paleozoya*, 1974; Gusev, 1975; Molostovskii, 1983; Ogg and Steiner, 1991; Lozovsky and Molostovsky, 1993). Of particular importance is the section of the Bol'shoe Bogdo Mt. in the Caspian area because of paleomagnetic determinations of the Bogdo Formation containing remains of protosuchians and ammonites of the *Tirolites cassianus* Zone within the N_2T interval.

The section of the Canadian Archipelago has 16 paleomagnetic zones of different ranks (Ogg and Steiner, 1991). The principal zones are six orthozones with normal and reversed polarity alternating in the same succession as in the similar stratigraphic interval of continental sections (Fig. 2).

In both sections, the lower Triassic column begins with a zone of normal magnetization (N_1T = GN_1). In the Russian plate, this column includes the larger part of the Vokhma Horizon; in the marine reference scale it comprises the lower half of the Griesbachian in the summary range of the *Otoceras concavum* and *O. boreale* zones (=*O.woodwardy* Zone of the general scale). This relationship of magnetozones corresponds well with the correlation made earlier for the lower Triassic horizons of the Moscow syneclise and Greenland on the basis of tetrapods and palynologic assemblages (Lozovskii, 1989).

The large interval of dominant reversed polarity, including two small zones with normal polarity (GR_1 + GR_2 + DR_1), is the most probable analogue of the R_1T zone of the continental section. Stratigraphically, it correlates with the upper half of the Griesbachian and with the complete range of the Dienerian.

The lower part of the next zone (N_2T) correlates with two N-zones of the Smithian Stage divided by the subzone of the reversed polarity.

Analogues of the upper part of the N_2T_1 zone in the Caspian area (Bol'shoe Bogdo Mt.) contain the protosuchians and ammonites of the *Tirolites cassianus* Zone, whereas in the stratotype of the Canadian Archipelago these analogues are missing due to erosion at the base of the Spathian Stage.

In the Mangyshlak Mountains, the upper part of the Dolnapa and the lower part of the Tartaly formations apparently correspond to the N_2T zone (Lozovskii, 1974).

The overlying R_2T_1 zone is an analogue of the SpR_1 zone in the lower part of the Spathian Stage. The same magnetization characterizes the upper beds of the *Tirolites* and *Columbites* zones of the Mangyshlak Mountains and synchronous upper Olenekian deposits of the Eastern Taimyr comprising the upper part of the *Dieneroceras* Zone and the *Olenecites pilaticus* Zone (*Paleomagnetizm paleozoya,* 1979; Gusev, 1975).

The youngest horizons of the continental Triassic (N_3T_1 zone), namely the deposits of the *Stacheites* Zone in Mangyshlak and the lower part of the *Keyserlengites subrobustus* Zone in the Tsvetkov Cape and Canada (SpN_1 zone), show normal polarity.

In the Canadian section, the paleomagnetic column is capped by a reversed polarity zone (SpR_2) confined to the upper half of the *Keyserlengites subrobustus* Zone. In European Russia (the Mangyshlak highlands) and Taimyr this part of the section is reduced because of the pre-Anisian erosion.

Distant correlations with the lower Triassic of the Tethys region are now possible owing to the results of paleomagnetic research in the Sichuan province of southern China, where the marine carbonate–terrigenous Feixianguan and Jialingjiang formations were studied in detail (Heller *et al.*, 1988; Steiner *et al.*, 1989). The data on conodont remains revealed here the Griesbachian, Dienerian (conditionally), Smithian and Spathian stages; the Triassic–Permian boundary is conformable without any apparent signs of erosion.

A summarized magnetostratigraphic section of southern China is structurally similar to sections of Canada and European Russia. This circumstance permits their zonal correlation with due regard for biostratigraphic data.

Most of the Griesbachian beds in both Sichuan and northern Canada show normal polarity and belong to the N_1T_1 zone. The next large R-interval (an analogue to the R_1T_1 zone) includes the Dienerian(?) and lower beds of the Smithian. Stratigraphically above them there is the aforementioned succession of zones N_2T_1, R_2T_1, and N_3T_1, which are identified as equivalents of the Spathian. The uppermost beds of the Spathian Stage correspond to an interval with reversed polarity; in the Canadian section, the zone SpR_2 (=the upper part of the *Keyserlengites subrobustus* Zone) is the most probable analogue of this interval.

CONCLUSION

The scale of magnetic polarity of the lower Triassic has been compiled during the last forty years, but only over the last few years has the work approached its goal. An important contribution has come from remote magnetostratigraphic correlations based on long-lasting and careful compilation of regional scales and accurate correlation of paleomagnetic and paleontological successions in marine and continental formations. The ranking of magnetic zones in accordance with their stratigraphic ranges and setting up the correlation framework on the basis of a few relatively large orthozones were of principal importance in this research.

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Paleoecosystem of *Dorsoplanites panderi* Phase in the Volga Shale-Generating Basin

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Abstract–The ecosystem of the formation period of low-calorie oil shales and sapropel clays at the final *Dorsoplanites panderi* phase of the basin evolution was examined. Paleobionomic reconstructions were made on the basis of investigations of fauna remains, and the analysis of the rock composition, fabrics, structures, and nannostructures, kerogen composition and properties, pelitic components, and other specific features that revealed the leading role of coccophore skeletal and organic remains in the composition of primary sediments. Sedimentation occurred under conditions of unstable salinity and a hydrodynamic regime within shallow (80–100 m maximum) and temperately warm waters. Reactions in the sediments of the shallowing basin altered from reducing to slightly oxidizing. Periodic breaks in anoxic environments within sediments, which were transformed into combustible shales with low heating efficiency, caused substantial degradation of the primary organic matter.

Key words: *paleoecosystem, low-calorie oil shales, sapropel clays, microfossils, the Volga region.*

Over a series of years, we investigated the stratigraphy of the shale-bearing sequence in the Volga region and the lithologo–geochemical peculiarities of their rocks in order to elaborate the legend for a large-scale mapping and to correlate sedimentary sections. This work enabled us to obtain data on the taxonomic composition of organic fossil remains, their ecologic and taphonomic peculiarities, and on the composition of the shale-bearing rock sequence that makes possible the consideration of this geological unit with reference to the respective ecosystem.

The shale-bearing formation proper (the central part of the Volga basin including the Central Volga region, left bank of Volga River, Obshchii Syrt, Northern Caspian region, and contiguous areas) is stratigraphically assigned to the *Dorsoplanites panderi* Zone of the middle Volgian substage. Its maximum thickness reaches 100–110 m. The formation comprises from five to six, to 12 strata of combustible shales displaying variable contents of organic matter (OM), and other compositional features related to formation pecularities of the rocks. (Kuleva *et al.*, 1992).

Using the ecosystem analysis, we studied dark-grey low-calorie oil shales and paragenetically associated black and dark-grey sapropel clays in the upper part of the shale-bearing sequence (beds with *Haplophragmoides volgensis inviolatus* Dain; Kuleva and Baryshnikova, 1988), where they are rich in organic fossil remains and constitute an essential part of the section.[1] In view of paleobionomic reconstructions, their paleobiofacies represent ecosystems of the lowest rank (Sobetskii, 1978) and are highly demonstrative for pertinent studies, as they bear valuable information about the organisms' life, relations and environments.

To determine the parameters of the basin ecosystem, the time of sediment accumulation, and starting materials for the above-mentioned rocks, we applied a complex of lithological geochemical and paleontological methods, and investigated organic matter, including the X-ray diffraction characteristics of the separated kerogen.

The study of paleontological remains was of primary importance for the paleobionomic reconstructions. In addition to the leading biota groups, all other diverse associations of organisms and preserved fragments of the ecological system were investigated.

The ecological–taphonomic study by the methods of Gekker (1955, 1966), Sobetskii (1978), Yanin (1983), and Morozova and Kuleva (1992), proved to be essential for the paleobionomic constructions. The method of actualism was widely used to reveal the ecological peculiarities of the fossils. Despite certain restrictions, this method greatly extends the opportunity to interpret the physical and geographical parameters of the environments: depth, hydrodynamics, temperature, and others.

The aforementioned shales and sapropel clays contain representatives of diverse faunal groups, which are predominantly pelecypods. Gastropods, chephalopods, and scaphopods are recorded in lesser amounts, while brachiopods and sea urchins are scarce. Among the microfauna, foraminifers are common and ostracods are occasional.

[1] According to the geological prospecting index, this is commercial Bed P of a complicated structure, up to 1.3 m thick.

Fig. 1. Oriented flaky–petaliform–coccolith nannostructure of low-calorie oil shales. Scanning electron microscope (SEM), Magnif. 7000.

In paleobionomic reconstructions, genus was the main operational unit, as it is generally accepted that only slight changes in ethology, feeding mode, and ways of obtaining food, as well as in organism reactions to abiotic factors is typical of this taxonomic rank. This is mostly relevant to pelecypods. Moreover, the generic taxa, unlike the species, are invariably understood by all specialists.

To obtain more reliable information on the microfauna biocoenoses in modern seas, it is important to determine the fauna abundance in a given sedimentation volume and to estimate quantitative propotions of taxa. We applied a uniform approach to studying fossil foraminifers. The total amount of forams in the complex was normalized to a common index: the "foraminiferal number" (FN), i.e., the number of shells per 100 g of dry sediment.[2] The foraminiferal number allows for greater objectivity in comparing different lithotypes and, thus, to obtain the most reliable information.

In addition to the general quantitative assessment, the numerical proportions of ecological foraminiferal types—secretory and sandy (agglutinating) benthic forms—were also revealed.

The complex lithological, geochemical and paleontological investigations of the rocks with regard to the aforementioned aspects, allowed us to characterize them rather comprehensively.

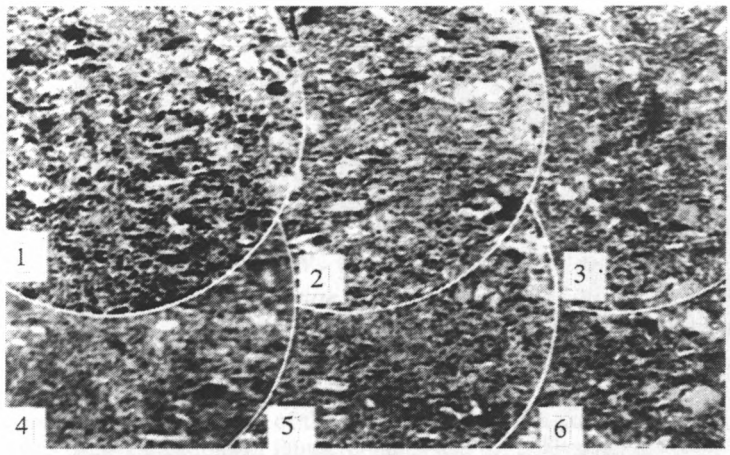
Shales are commonly dark gray, in some cases even black, occasionally with a brownish shade. They are variably calcareous, marked by an elevated content of aleuritic particles (up to 30%), and by the presence of macro- and microlenses of carbonate-free clays. Their horizontal laminated structure is formed by an irregular distribution of organic matter in layers and an insignificant variations in the granulometric composition. In some intervals, this structure is characterized by the occurrence of accumulated organic remains. All of these factors are responsible for the thin-platy and laminated structure of the rocks. Moreover, the lamination of low-calorie shales may have resulted from diagenetic processes.

The microstructure of shales is microlenticular and coagulative; the texture is aleuro–pelitic; and the nannostructure is oriented, flocculent–petaled–coccolithic (Fig. 1). Commonly, coccolithes are poorly preserved, particularly in lenses of carbonate-free clays, which display a relict coccolithe nannotexture (Fig. 2); the latter suggests the diagenetic dissolution of sapropel ooze carbonates.

The X-ray diffraction analysis showed that the kerogen extracted from shales has a sapropel composition with an insignificant amount of humic components.

Judging by the OM character (its amount ranging from 13 to 24%), the shales are of the alginite–sapropel–collinite type. The OM composition is dominated by oxidized (brown) sapropel microcomponents with an admixture of allochthonous coalified plant remains and secondary humic acids, formed while alginite was disintegrated by aerobes during sedimentation and by anaerobes in the course of diagenesis. This is a result of the nonequilibrated state of the sedimentation medium, and eH, and pH fluctuations in the basin during the period under consideration.

[2] According to Fursenko (1978) and Grigyalis (1985), the complex composed of the total collection of forams buried in the sediments that were identified in the rock.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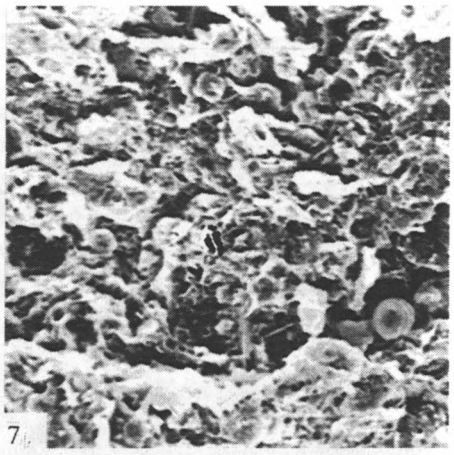
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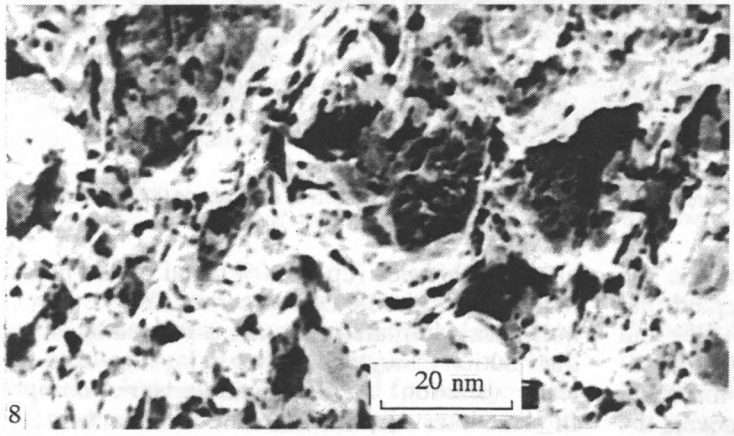
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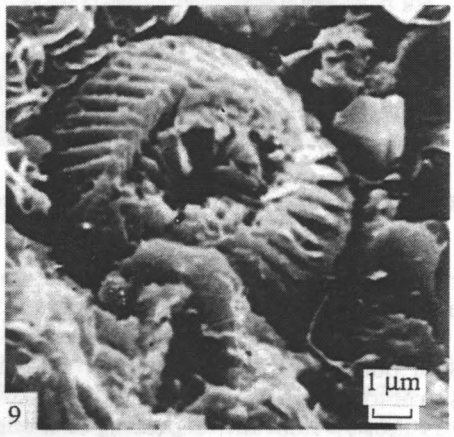
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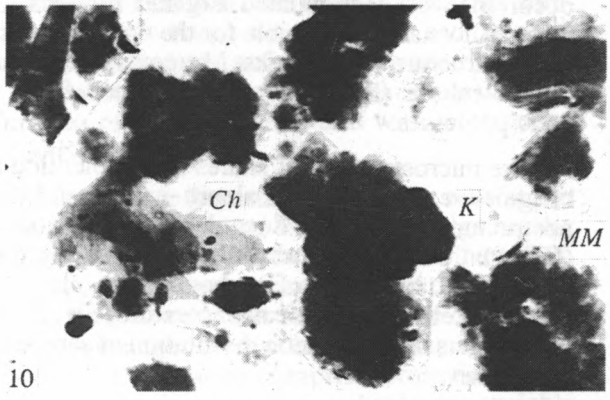
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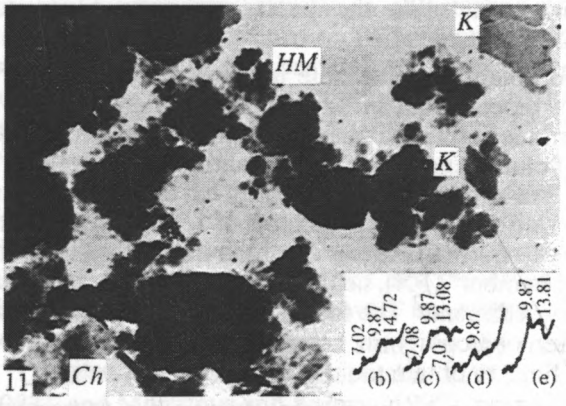
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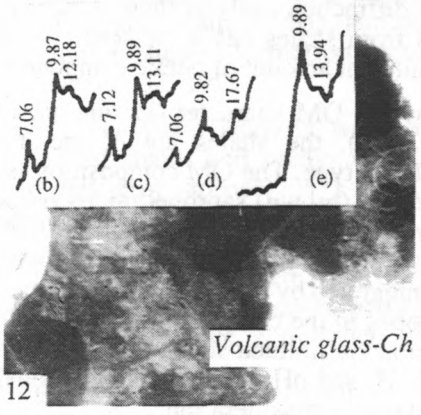
11
Ch


7.06
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Volcanic glass-Ch
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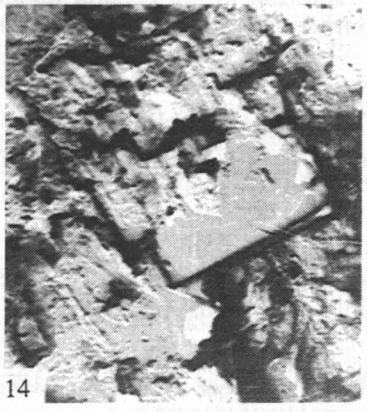
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The ash content in the investigated oil shales is up to 60–65%. This is evidently a result of an increased influx of terrigenous material into the basin. Particularly high is the content of clay minerals, which, along with organo-mineral aggregates, form separate microlenses with similar optical orientations of particles. Hydromica (40–60%), montmorillonite (5–25%), chlorite (upto 5%), kaolinite (5–10%), and admixtures of quartz, feldspar, and siderite were identified in the pelitic fraction by the X-ray diffraction analysis. The diffractograms of fractions ~1 μm with characteristic reflexes of these clay minerals are given in Figs. 2 and 3.

In addition to the mentioned minerals, the electron microphotographs revealed palygorskite and halloysite. Moreover, the aleuritic fraction contains minerals labile under weathering, among them epidote, hornblende, and biotite; it also shows a specific enrichment in lithophile trace elements: Cr, V, and Ti. These rock components and palygorskite are similar to the arid lithogenesis products recorded in sandy–silty–clayey rocks of the Upper Permian–Triassic provenance (Yanochkina, 1966). Allogenic kaolinite, halloysite, and several other clay minerals in the shales might have been derived from the mid-Late Triassic–Early Jurassic humid weathering crust of neighboring areas. The composition of these minerals was investigated by Gutsaki and Gudoshnikov (1969).

A characteristic feature of dark-gray shales is the alteration of their clay particles at the diagenetic stage, which was expressed in the defects of hydromica and chlorite structures, dissolving of carbonates, and replacement of coccolithes by authigenic clay minerals, including kaolinite and halloysite. These processes were evidently a result of partial basin freshening accompanied by the pH reduction in the bottom water layer (Pugachev and Kalmykova, 1985).

Yudovich and Ketris (1988) reported that shales with siderite—the product of gley carbonic acid diagenesis—might be formed in a freshened marine environment, when all carbonates become unstable (Curtis, 1980). Analogous phenomena are recorded in the Jurassic oil shale of the Pechora River basin (Romanovich, 1981), and in the Volga basin (Pugachev *et al.*, 1985).

Pugachev believes that in such a case the sediments are under weak oxidizing rather than reducing conditions. This was confirmed by the analysis of the chemical composition of residual solutions in clays and oil shales in Bed P. Romanovich regards repeated water freshenings as catastrophic for marine fauna and the main cause of the explosion in the abundance of primitive shale-generating organisms in stress conditions in ecosystems.

Sapropel clays associated with the shales under consideration are dark gray, aleuritic, and variably calcareous. By their lithological and geochemical features, the clays resemble the shales, differing mainly in their reduced OM content. Their formation was evidently accompanied by a more intensive influx of terrigenous material and a concurrent decrease in phytoplankton productivity which resulted from the temporary normalization of the conditions favorable for the active development of typical marine faunal groups. Their remains are commonly more abundant in clays than in shales.

The interpretation of the litholgical and geochemical features of rocks enabled us to make the following conclusions pertaining to the parameters of the sedimentation: biogenic components of nannoplankton (coccolithophores) of both the gametophyte and sporophyte types with different generations rates, were the main initial components of OM during the accumulation of muddy sapropel sediments (Bukina, 1988). According to the evidence presented by Egorov (1985), and Gorlenko and Zhmur (1989), OM in shales is mostly composed of the remains of blue-green algae (*Cyanophyceae*). They may have been in symbiosis with coccolithophores or alternatively, developed over sediments rich in skeletons of these algae, supplementing coccolithophore "mats" and uniting them into geological bodies.

The character of rock structures corresponds to temperate quiet hydrodynamic conditions with some elements of activation, while the geochemical features of the rocks indicate a certain decrease in the normal marine salinity in geological "moments" when the original sediments were accumulated.

Oryctocoenoses in the rocks under consideration are rich in diverse organic remains. These are bivalves *Phacoides, Oxytoma, Astarte, Buchia, Entolium, Inoceramus*, scaphopods *Dentalium*, gastropods *Scurria*, brachiopods *Lingula*, ammonites, small fish bones and scales, foraminifers, and coccolithes. Traces of deposit feeders, which are observed quite frequently, is a characteristic element of the assemblages.

The bivalves are predominant, while small scaphopods (*Dentalium*) are also abundant, and brachiopods are rare. There are many ammonites, and their taxonomic composition is typical for the *Dorsoplanites panderi* Zone. Apart from the zonal genus, there are *Zaraiskites* and *Pavlovia*. Remains are commonly small in size, not more than 0.5 cm; only certain scur-

Fig. 2. Characteristic features of decarbonized sapropel clays.

1–6: Pelitic and aleurolite–pelitic structure of clays. Polarizing microscope (PEM), magnif. 1000; 1, 3, 5: crossed polaroids; 2, 4, 6: parallel polaroids; 7–9: flaky and relict coccolith nannotexture of clays, SEM, magnif. 1000–7000; 10–14: Morphology of pelitic particles, observed in suspension. PEM, magnif. 5000–13000; (*HM*) hydromica, (*MM*) montmorillonite, (*K*) kaolinite, (*Ch*) chlorite. X-ray diffraction characteristic of the pelitic fraction: (b) natural oriented sample; (c) treated in HCl; (d) saturated with glycerin, (e) calcinated at 600°C during 1 h.

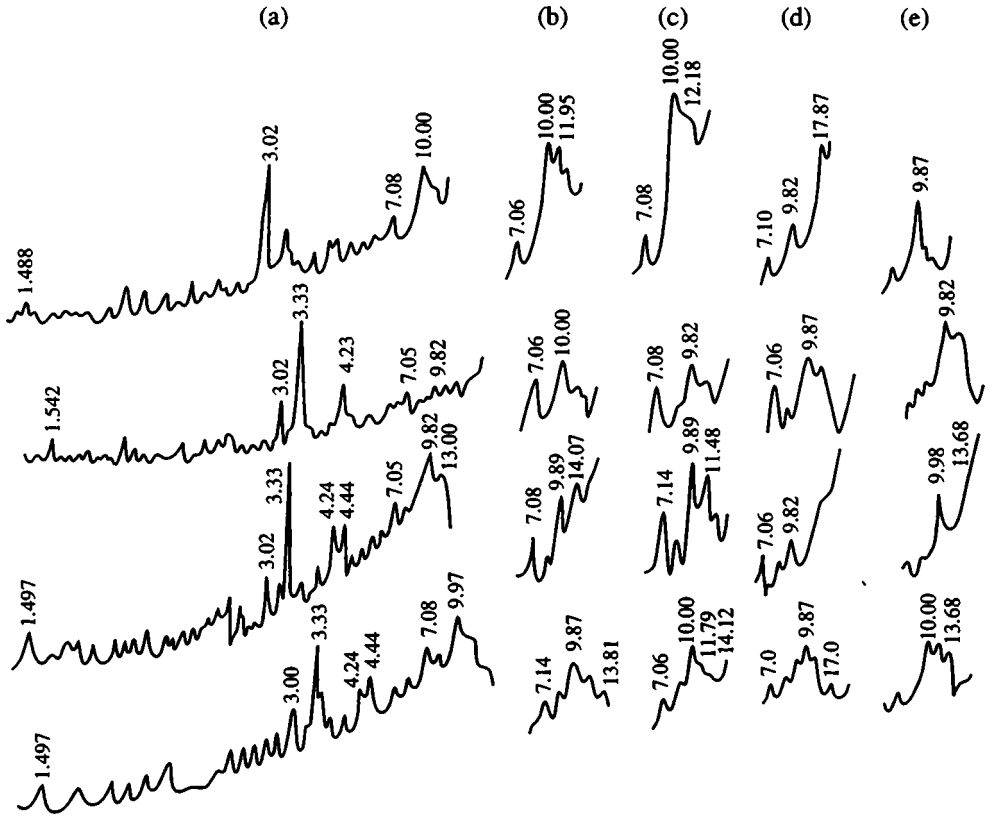

Fig. 3. Difractogram of fractions <1 μm; (a) natural non-oriented sample; (b) natural oriented sample; (c) treated in HCl; (d) saturated with glycerin; (e) calcined at 600°C during one hour.

rias, isolated buchias, inocerams and ammonites are large.

Many faunal remains are poorly preserved; in most cases, nothing but a thin calcareous membrane is left from the shells. No traces of their displacement are observed: shells are not damaged, the valves of bivalves might be open, but are not dismembered. Regular lit-par-lit taphonomic burials suggest that they were formed under conditions of moderate water movements, accompanied by bottom currents. Kuznetsov (1980) suggests that such conditions existed in the upper part of the middle sublittoral. Some platy jointing planes contain mass accumulations of young ammonites, bivalves and tusk shells, exemplifying the lit-par-lit type of concentrated burials (Sobetskii, 1978).

Abundant representatives of *Phacoides* in oryctocoenoses and the lithological composition enabled us to unite the considered rocks into the *Phacoides* paleofacies of the silty–clayey sublittoral (Kuznetsov, 1980).

The taphonomic peculiarities show that given oryctocoenoses represent the autochthonous burials and main components of paleobiocoenoses; the taxonomy of their representatives can be described by means of the ecological parameters. The results of the ecological interpretation are summarized in the table based on generic taxa.

The FN in foraminiferal oryctocoenoses of sapropel clays ranges from 150 to 300 specimens; the species composition is poor, with no more than 15 species. The specific feature of the complexes is the predominance of secretory forms. These are representatives of *Lenticulina, Marginulina, Citharina*, and *Saracenaria* genera, which have a massive porcelain-like shell with diverse and well-developed structures (ribs, keels, umbilical callus); *Lenticullina* forms are the most numerous. Agglutinating specimens (*Haplophragmoides*) make up 2–3%.

The predominance of secretory forms and their large shells indicate the normal salinity in the basin and its shallow depth.

In the low-calorie oil shale, with approximately the same FN but less diverse species composition (no more than ten species), the proportion of ecological groups is quite different: the abundance of agglutinating species ranges here from 75 to 97%. These are mostly representatives of the *Haplophragmoides* and *Ammobaculites* genera. Secretory forms are rare, likely due to their solution in the sediment.

The ecological information obtained from macro- and microfossils allows us to generally characterize the biota at the time of sapropel ooze formation in the following aspects.

Ecological characteristics of bivalves, gastropods, scaphopods, and brachiopods of the *Dorsoplanites panderi* phase in the Volga Basin

Fauna	Ecological factors																
	ethological			trophic		edaphic		hydrodynamic		bathymetric		oxide		thermal			salinity	
	sessile	infaunal	vagile	sestonophags (filter feeders)	phytophags	eurytopic	sandy muds	weak	moderate	eurybathic	shallow water	euryoxibionic	stenooxibionic	cryophilic	eurythermal	temperate–warm waters	stenohaline (without deviations)	stenohaline (common), though resistant against freshening
Phacoides		+		+		+		+			+		+			+		+
Astarte	+			+			+		+	+			+	+			+	
Oxytoma	+			+		+		+			+		+			+	+	
Buchia	+			+		+			+	+			+	+			+	
Entolium			+	+			+	+		+		+				+	+	
Inoceramus	+			+		+			+	+			+	+			+	
Scurria			+		+		+	+			+		+			+	+	
Dentalium		+		+			+	+		+			+		+		+	
Lingula		+		+			+	+			+		+		+			+

Taxonomic aspect. The basin was inhabited by representatives of different marine fauna and flora, comprising unicellular to highly evolved species. Macrobenthos was dominated by bivalves; scaphopods and gastropods (*Scurria* forms) were abundant; brachiopods, diverse limivorous forms, and benthic macrophytes were also present. Benthic microfauna was represented by forminiferal assemblages of both agglutinating and secretory types. The pelagic zone was inhabited by ammonites, fish, and, probably, by marine saurs, as well as by phytoplankton, including coccolithophores and radiolarians.

Ethologic aspect. Infauna representatives—bivalves *Phacoides* and scaphopods (*Dentalium*)—were most widespread. Among epifauna, there were gastropods *Scurria* in symbiosis with algae. The amount of other benthic forms—*Astarte* with byssus, *Buchia*, *Inoceramus*, and infaunal *Lingula*—were insignificant.

Trophic aspect. Apart from phytophagous *Scurria* and sediment-inhabiting detritophags, all other epi- and infaunal representatives were vagile or sessile sestonophags.

Thermal aspect. Judging by the recent distribution of genera, moderately thermophilic bivalves *Phacoides* and gastropods *Scurria* prevailed in biocenoses. However, some cryophilic forms such as *Buchia, Inoceramus*, and *Astarte* along with *Oxytoma* and Pectinacea inhabiting temperate water were also identified. The ammonite species, remains of which were found in the rocks, were widespread in boreal seas.

Salinity aspect. Ammonites and the majority of other mollusks and foraminifers were marine stenohaline organisms. Only *Phacoides* and *Lingula* were able to resist a decrease in the marine salinity.

Hydrodynamic, bathymetric, and edaphic aspects. Most representatives of the benthic filter feeders preferred relatively shallow depths with sandy to muddy grounds, and weak hydrodynamic activity. The disribution of phytophagous *Scurria* was confined to the euphotic zone (down to the depth of 80–100 m).

The ecosystem parameters at the time of accumulation of the original sediments now transformed into low-calorie oil shales and sapropel clays, were certainly influenced by the general character of the sedimentation cycle confined to the upper part of the *Panderi* zone, namely to beds with *Haplophragmoides volgensis inviolatus* Dain and *Marginulina formosa* Mjatl. This transgressive-regressive cycle was accompanied by the basin widening and shallowing against the background of the general elevation of the basin floor and adjacent land areas (Bukina, 1988; Kuleva, 1992). At that time, the humid Upper Triassic–Lower Jurassic and arid Permian–Lower Triassic deposits were synchronously exposed to the day surface, and served as a provenance for sediments of the *Panderi* phase. A slightly elevated erosion basis caused an influx of mostly dispersed material into the basin, while in periods of relatively great uplift, it was represented by coarser aleuritic particles. A good preservation of the Upper Permian unstable minerals in the rocks suggests the general climate aridization. However, certain climate humidization occurred at the "moments" of sapropel ooze formation, and was apparently caused by more free connections with the oceanic boreal basin that resulted in a great number of boreal forms among invertebrates, and by heavy rains, which freshened surface waters of the marine basin.

The considered rocks are attributed to the bionomic environment of the *Phacoides* paleofacies produced by the following factors.

Bottom plants, which served as food for numerous *Scurria*, suggest a depth interval above 80–100 m, and, consequently, sufficient illumination. The predominance of sestonophags (mainly *Phacoides*) was associated with an abundance of suspended food, apparently due to the mobility of the water mass in the currents, which supplied food and ensured a normal gas regime for numerous epi- and infaunal forms both in the bottom waters and in upper sedimentary layers. Moreover, the OM sorbed on clayey particles of the sediment may have also been used as food.

Coccolithophores occupied a special place among the inhabitants of the basin: their high productivity promoted the formation of sapropel ooze. The extreme development of coccolithophores can be associated with the appearance of stress situations, caused by appreciable water freshening (Romanov, 1981) that was ruinous for stenohaline marine fauna. This is evident from the presence of concentrated taphonomic burial types in shales and sapropel clays. The saturation of water with uranium may also be a factor that intensified the stress situation and stimulated the extreme phytoplankton productivity (Neruchev, 1982).

Thus, macro- and microalgae were the principal producers of the primary OM in the basin, and microzooplankton (mollusk larvae, radiolarians) and benthos were the primary consumers. In turn, benthos feeding on microzooplankton served as food for cephalopods. The sediment, rich in OM, was reworked by nonassorting detritophags, the mud feeders and bacteria which were active reducers of the ecosystem. Our reconstruction of the ancient ecosystem is based on detailed research, which allowed us to trace the evolution of the *Panderi* phase and to supplement the incomplete information on ecosystem relations, reflected in the geological record. Such investigations contribute to clarifying the general development of ancient basins.

Reviewer I.A. Basov

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Stratigraphy and Radiolarians from the Type Section of Perapedhi Formation (Upper Cretaceous of Cyprus)

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Abstract—Late Santonian–Early Campanian radiolarians were found in the Perapedhi Formation type section (Upper Cretaceous, Cyprus). The fossils allow us to conclude that the development of the underlying Troodos Ophiolite Complex was completed at the end of the Santonian–beginning of Campanian. Five new species of Radiolaria were described: *Archaeospongoprunum distributum*, *Triactoma fragilis*, *Stichomitra perapedhia*, *Theocampe*(?) *cypraea*, and *Theocoronium subtriquetrus*.

Key words: *Upper Cretaceous, Late Santonian, Early Campanian, stratigraphy, radiolarians, Cyprus.*

The Perapedhi Formation represents the most ancient deposits of the Troodos ophiolite complex sedimentary cover, which directly overlie upper pillow lavas or associated volcaniclastic breccias. Stratigraphic information from the Perapedhi Formation is of primary importance for the reconstruction of the geologic history of Cyprus in the Late Cretaceous and for structural analysis of this area, because the deposition of the Perapedhi Formation took place directly after the formation of the Troodos ophiolites and before their collision with the allochthonous Mamonia Complex (Robertson and Woodcock, 1979). Indeed, the age of Perapedhi Formation is questionable. Some authors suppose it to be the Campanian (Swarbrick and Robertson, 1980), or Turonian (Blome and Irwin, 1985). Thurow (1991) concluded that it spans a wide stratigraphic range—from the Turonian to Lower Campanian. Paleontologic remnants from the Perapedhi Formation are represented only by radiolarians, which have not been studied in detail.

The Perapedhi Formation type section is located in the central part of Cyprus near the Perapedhi Village (Fig. 1). It was studied and sampled by an expedition of Geological Institute of the Russian Academy of Sciences in 1990. Outcrops of the Perapedhi Formation were studied in a roadcut near the Saittas Village (150 m from the Perapedhi–Saittas–Kouka road connection). The sequence is represented by (Fig. 1):

1. Volcaniclastic breccias with a tuffaceous matrix and fragments of basalt and diabase. The observed thickness in this outcrop is 1 m. The same breccias have thicknesses of several dozen meters south of the roadcut.

2. Red and yellow to rust-colored clays, which fill the pockets on the eroded surface of breccias. The thickness varies from 0 to 1 m.

3. Black and dark-brown loose manganese-bearing mudstones with weakly rounded fragments (up to 2 cm) of yellowish-gray phosphatic clayey limestones. Pebbles of limestones are parallel to bedding, and contain poorly preserved radiolarians (probably Late Cretaceous). The thickness is 0.1–0.2 m.

4. Umbers that are dark brown and yellow, platy, sometimes massive, with brownish-black intercalations, massive, Fe–Mn-cherts, rare intercalations (1–2 cm) of yellow tripoli-like silty radiolarites, a few intercalations (1 cm) of pinkish-white carbonatized radiolarites, and rare iron-manganese concretions. Radiolarites were found in the upper part of this member. They contain the abundant radiolarian assemblage: *Acaeniotyle* spp. (Empson-Morin), *Actinomma* sp., *Alievium gallowayi* (White), *Alievium superbum* (Squinabol), *A.* sp., *Archaeospongoprunum bipartitum* Pessagno, *A. distributum* Bragina, sp. nov., *Crucella cachensis* Pessagno, *C. messinae* Pessagno, *Halesium* sp. [Blome and Irwin, 1985], *Multastrum regalis* Vishnevskaya, *Orbiculiforma vacaensis* Pessagno, *Patellula verteroensis* (Pessagno), *Patulibracchium* sp. aff. *P. dickinsoni* Pessagno, *P.* spp. aff. *P. lawsoni* Pessagno, *P.* sp., *Phaseliforma carinata* Pessagno, *Praeconocaryomma irwini* Pessagno, *P.* sp. cf. *P. lipmanae* Pessagno, *P. universa* Pessagno, *Pseudoaulophacus floresensis* Pessagno, *P. lenticulatus* (White), *P. praefloresensis* Pessagno, *Spongosaturnalis* (?) *ichikawaji* Foreman, *S.* sp. aff. *S. horridus* (Squinabol), *S.* sp. ex gr. *S. hueyi* Pessagno, *S.* sp., *Stylosphaera pusilla* Campbell et Clark, *Triactoma fragilis* Bragina, sp. nov., *Afens lirioides* Riedel et Sanfilippo, *Amphipyndax pseudoconulus* (Pessagno), *A. stocki* (Campbell et Clark), *Annikaella omanensis* De Wever, *Archaeodictyomitra regina* (Campbell et Clark), *Artostrobium urna* Foreman, *Bathropyramis* sp., *Cornutella californica* Campbell et Clark, *C.* sp., *Cryptamphorella conara* (Foreman), *Dictyomitra* sp.

cf. *D. andersoni* (Campbell et Clark), *D. densicostata* Pessagno, *D. formosa* Foreman, *D. multicostata* Zittel, *D.* sp. ex gr. *D. torquata* Foreman, *Neosciadiocapsa jenkinsi* Pessagno, *Pseudodictyomitra* sp. aff. *P. nakasekoi* Taketani, *Sciadiocapsa cansia* Foreman, *S.* sp., *Stichomitra cechena* Foreman, *S. perapedhia* Bragina, sp. nov., *S.* sp. aff. *S. insignis* (Squinabol), *Squinabollum fossilis* (Squinabol), *S.* sp. A, *S.* sp. B, *Theocampe* (?) *cypraea* Bragina, sp. nov., *Theocampe* sp. cf. *T. salillum* Foreman, *T.* sp., *Theocapsomma comys* Foreman, *Theocoronium subtriquetrus* Bragina, sp. nov., *Theocorys antiqua* (Squinabol), and *Xitus asymbatos* (Foreman). The supposed age of this assemblage is Late Santonian–Early Campanian (see Plates). The thickness is 10 m.

Tectonic contact (fault) separates the following beds:

5. Greenish grey and brownish grey silty clays, intercalated with light-grey opoka-like siltstones with the radiolarians: *Afens lirioides* Riedel et Sanfilippo, *Alievium gallowayi* (White), *Clathropyrgus* sp., *Crucella robusta* Bragina, *Dictyomitra densicostata* Pessagno, *D. formosa* Foreman, *D.* sp. ex gr. *D. torquata* Foreman, *Patellula planoconvexa* (Pessagno), *Praeconocaryomma ... ersa* Pessagno, *Pseudoaulophacus floresensis* Pessa... *P. praefloresensis* Pessagno, *Rhopalastrum* sp. cf. ... *tumidum* Lipman, *Amphipyndax pseudoconulus* (Pessagno), *Stichomitra* sp. cf. *S. campi* (Campbell et Clark), and *Xitus asymbatos* Foreman, presumably implying the Campanian age. This unit is an equivalent of the Kannaviou Formation (Upper Campanian–Lower Maastrichtian), which overlies the Perapedhi Formation in most sections (Swarbrick and Robertson, 1980). The observed thickness is 12 m.

The radiolarian assemblage from layer 4 is extremely rich and taxonomically diverse (see Plates). Present here are abundant and massive thick-walled forms, such as *Alievium gallowayi* and *Stichomitra*, and very fine forms, such as *Spongosaturnalis* and *Neosciadiocapsa*. The assemblage also includes specimens with pseudoaulophacoid, cancellate, and a coarse reticulate structured skeletal wall. It should be mentioned that this assemblage was extracted from cherts by means of hydrofluoric acid. Large specimens with thin fragile test usually cannot be extracted by this technique. Nevertheless, such morphotypes are widely represented in this assemblage. The good preservation of such fragile specimens shows that nearly all taxa of the assemblage were extracted and studied in details.

The assemblage shows a high taxonomic diversity of such genera as *Dictyomitra, Stichomitra, Spongosaturnalis, Crucella*, and *Praeconocaryomma*. It is difficult to suggest the quantitative domination of any species. Most common are: *Annikaella omanensis, Dictyomitra densicostata, Cryptamphorella conara, Patulibracchium californiaensis, Pseudodictyomitra* sp. aff. *P. nakasekoi, Spongosaturnalis* (?) *ichikawayi, Stichomitra perapedhia*, and *Triactoma fragilis*. On the other hand, *Afens lirioides, Artostrobium urna*, and *Pseudoaulophacus lenticulatus* are relatively rare. Many species of this assemblage are extinct in the Campanian, including *Alievium superbum, Archaeospongoprunum bipartitum, Theocampe urna, Crucella messinae*, and *C. cachensis*. This indicates that the age of the assemblage cannot be younger then the Campanian. In addition, *Sciadiocapsa cansia* and *Theocapsomma comys* first appear in the Campanian. *Alievium gallowayi, Patellula verteroensis, Pseudoaulophacus floresensis, P. lenticulatus, Afens lirioides, Amphipyndax pseudoconulus, Artostrobium urna, Cryptamphorella conara*, and *Xitus asymbatos* are typical of the *Amphipyndax pseudoconulus* Zone of the tropical Campanian (Empson-Morin, 1981).

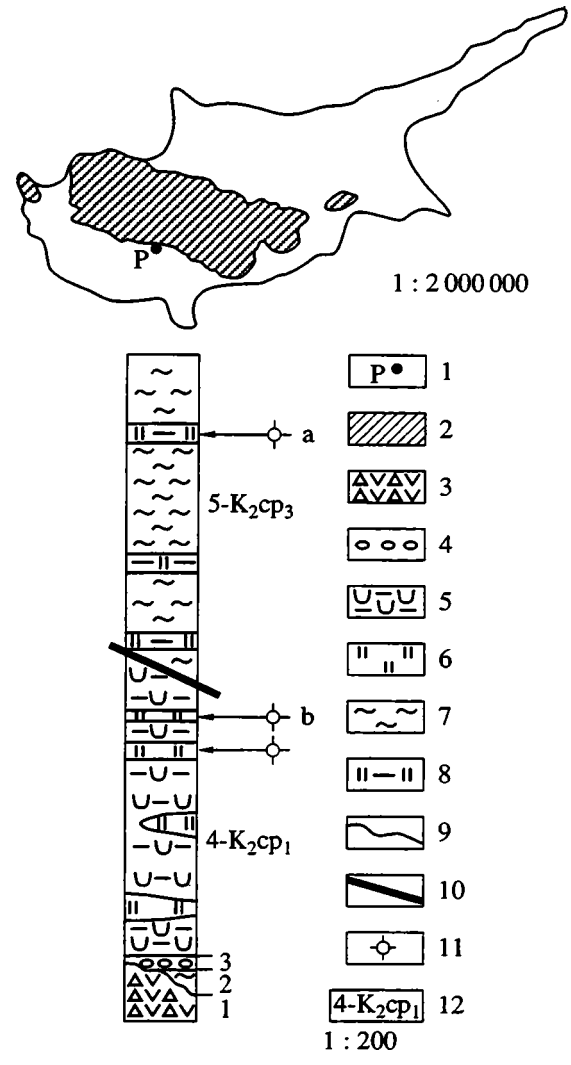


Fig. 1. Type Section of the Perapedhi Formation.
(1) locality of section; (2) outcrops of the Troodos ophiolite complex; (3) volcaniclastic breccias; (4) conglomerates; (5) umbers; (6) cherts; (7) clays; (8) opoka-like siltstones; (9) erosion level; (10) tectonic contact; (11) Campanian (a), and Santonian to Lower Campanian (b) radiolarians; (12) number of bed and its age.

Nevertheless, some features of this assemblage allow us to suppose the Late Santonian–Early Campanian age of the studied deposits. It should be noted first that *Theocorys antiqua* and *Squinabollum fossilis* are extinct at the Santonian–Campanian boundary. Second, the relatively rare tests of *Amphipyndax pseudoconulus*, which are represented by specimens with a weakly developed nodose wall structure, were recovered together with the typical specimens.

Many species of this assemblage are similar to those from the coeval Californian assemblages. Approximately half of the taxonomic composition is represented by species known from the Great Valley Sequence (Pessagno, 1976), and approximately 25% are typical of the Coast Range formations (Foreman, 1968). On the other hand, the taxonomical affinity with tropical oceanic assemblages is not so high (Foreman, 1975; Empson-Morin, 1981, 1984; Sanfilippo and Riedel, 1985). This can probably be interpreted as a result of the partial, incomplete study of oceanic assemblages. The radiolarian biostratigraphy of the oceanic Cretaceous is based on a limited amount of well-studied taxa, which were used as key groups. This resulted in the relatively low detalization of the radiolarian biostratigraphy of the uppermost Cretaceous deposits. For example, the Campanian and Maastrichtian were subdivided into two radiolarian zones, while the subdivision of the Upper Jurassic and Lower Cretaceous stages was more detailed: each of them was discriminated into several zones (Baumgartner, 1984). A complete study of the taxonomic composition of all the assemblages is necessary for further progress in the radiolarian biostratigraphy of the Upper Cretaceous.

The Perapedhi Formation in the type section is the Late Santonian–Early Campanian in age. Previous investigators also found Turonian radiolarians from the Perapedhi Formation in the Mangaleni section southeast of the Troodos Massif (Blome and Irwin, 1985; Thurow, 1991). This age discordance requires an explanation. It may be caused by the diachronous character of the lower boundary of the Perapedhi Formation. A more ancient radiolarian assemblage could also be extracted from the pebbles of the basal horizon of the Perapedhi Formation in the type section. On the other hand, the Perapedhi Formation overlies volcaniclastic breccias in the type section, and directly lies on the upper pillow lavas in the other sections. The age of volcaniclastic breccias from the Perapedhi Section can differ from the age of pillow lavas. These breccias were insufficiently studied and could be of the Turonian to Santonian age. Despite many unsolved problems, the obtained data show that the development of the Troodos Ophiolite Complex was completed before the end of the Santonian or beginning of the Campanian.

Paleontologic descriptions

Order Spumellaria Ehrenberg, 1875

Superfamily Spongodiscoidea Haeckel, 1862

Family Sponguridae Haeckel, 1862

Genus Archaeospongoprunum Pessagno, 1973

Archaeospongoprunum distributum Bragina, sp. nov.

Plate I, fig. 7.

Etymology. Distributum Lat.—divided.

Holotype. GIN 4840-1, Cyprus, Perapedhi Village, Upper Santonian–Lower Campanian, Perapedhi Formation.

Description. Archaeospongoprunum has massive and edged spines, gradually thinning toward the distal parts. The termination of one spine can be slightly spiral. The test is elongated with clear median incision, and has the shape of two small balls with mutual axis. The test walls have circular pores in hexagonal and pentagonal pore frames with small sharp horns at vertices. There are 9 to 12 pores on the line connecting the basal parts of the main spines. Sometimes the wall structure is similar to the pseudoaulophacoid one. The length of the spines is similar to the length of the test (without spines).

Dimensions. Length of spines is 90 μm; maximum length of test is 120 μm (without spines).

Comparison. This species differs from the most similar species *A. bipartitum* Pessagno by the shorter test with wall structure of the pseudoaulophacoid type, by more massive and short spines, and by a clearer median incision.

Distribution. Upper Cretaceous, Upper Santonian–Lower Campanian, Perapedhi Formation, Cyprus.

Material. 9 specimens from type locality.

Superfamily Actinommoidea Haeckel, 1881

Family Xiphostylidae Haeckel, 1881

Genus Triactoma Rust, 1885

Triactoma fragilis Bragina, sp. nov.

Plate I, fig. 11.

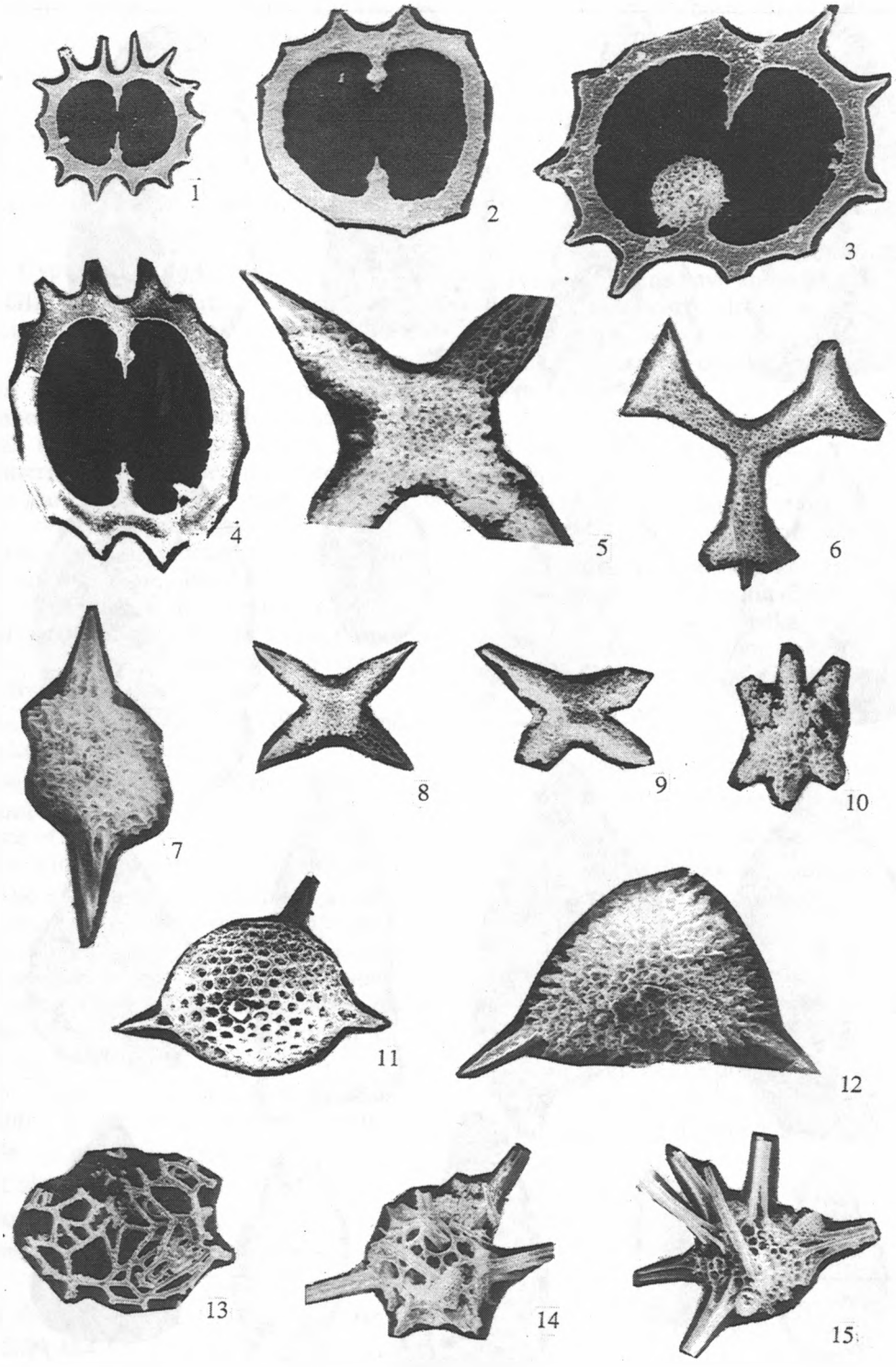
Etymology. Fragilis Lat.—fragile.

Holotype. GIN 4840-2, Cyprus, Perapedhi Village, Upper Santonian–Lower Campanian, Perapedhi Formation.

Description. The test is spherical with a thin reticulate wall and circular pores. Pore frames are hexagonal and pentagonal with small smooth nodes at vertices. There are 10 to 12 pore frames on the axial line prolongation of the main spines. Spines are short and very thin; their length is equal to a quarter of test diameter. Three large elliptical pores are visible near the base of each spine. On the spine, they pass into small grooves, which terminate at the middle part of the spine. Spines quickly become thinner toward the ends.

Dimensions. Diameter of test is 120 μm, and the maximum length of spines is 40 μm.

Comparison. The new species differs from *T. paramexicana* Pessagno by the smoother surface of test, shorter spines, and by larger pores, especially near the base of the main spines.



Radiolaria from the Type Section of Perapedhi Formation (Plates I–III).

Plate I.

(1) *Spongosaturnalis* sp. aff. *S. horridus* Squinabol, ×100; (2, 3) *Spongosaturnalis* sp., ×200; (4) *Spongosaturnalis* sp. ex gr. *S. hueyi* Pessagno, ×200; (5, 8) *Crucella cachensis* Pessagno, (5) ×200; (8) ×100; (6) *Patulibracchium* sp. aff. *P. lawsoni* Pessagno, ×200; (7) *Archaeospongoprunum distributum* Bragina, sp. nov., ×200; (9) *Crucella messinae* Pessagno, ×100; (10) *Multastrum regalis* Vishnevskaya, ×100; (11) *Triactoma fragilis* Bragina, sp. nov., ×200; (12) *Alievium* sp., ×200; (13–15) *Actinomma* sp., ×100.

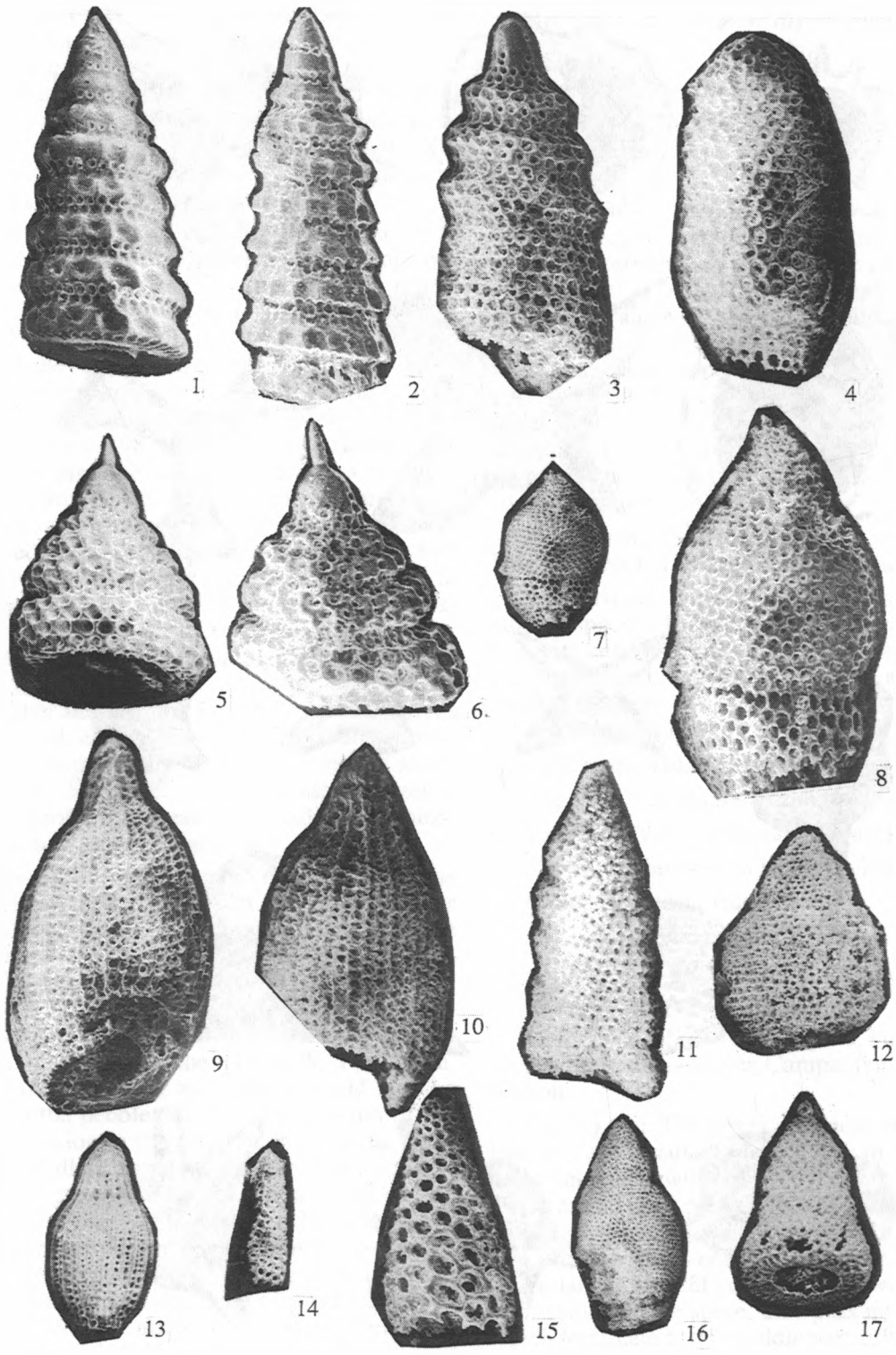


Plate II.

(1, 2) *Pseudodictyomitra* sp aff. *P. nakasekoi* Taketani, ×200; (3) *Stichomitra* sp. aff. *S. insignis* (Squinabol), ×200; (4) *Theocampe*(?) *cypraea* Bragina, sp. nov., ×200; (5, 6) *Stichomitra perapedhia* Bragina, sp. nov., ×200; (7, 8) *Theoperidae* gen. indet.: (7) ×100; (8) ×200; (9, 10) *Stichomitra* (?) sp., ×200; (11) *Stichomitra cechena* Foreman, ×200; (12, 16) *Theoperidae* gen. indet., ×150; (13) *Theocampe*(?) sp., ×100; (14, 15) *Cornutella* sp.: (14) ×100; (15) ×200; (17) *Stichomitra* sp. B, ×100.

Distribution. Upper Cretaceous, Upper Santonian–Lower Campanian, Perapedhi Formation, Cyprus.

Material. Eight specimens from the type locality.

Order Nassellaria Ehrenberg, 1847

Superfamily Eucyrtidioidea Ehrenberg, 1847

Family Eucyrtidiinae Ehrenberg, 1847

Genus Theocampe Haeckel, 1887

Theocampe (?) *cypraea* Bragina, sp. nov.

Plate II, fig. 4.

Etymology. Cypraea—from Cyprus.

Holotype. GIN 4840-3, Cyprus, Perapedhi Village, Upper Santonian–Lower Campanian, Perapedhi Formation.

Description. The test is spindle-shaped and consists of eight to nine chambers. Cephalis is subspherical with few small and sparsely distributed pores. Postcephalic chambers have large, identical pores arranged in three to five transversal rows in each chamber. The lengths of the chambers are slowly increasing; the widths of the chambers are increasing up to the fourth chamber, and the next three chambers are constant in this parameter. The width of last chamber is decreasing with the thinning of test wall. The terminal part of the test has an incomplete shape with uneven edges. The chambers are divided by inner ridges.

Dimensions. Length of the test is 300 μm, and the maximum thickness of the test is 150 μm.

Comparison. The new species differs from *T. altamontensis* (Campbell et Clark) by the smooth surface of the test, absence of grooves with pore rows, and by the absence of an incision between the thorax and abdomen.

Remarks. The new species is debatably assigned to the genus *Theocampe*. Its outer shape and, especially, shape of the cephalis and thorax show considerable affinity with representatives of this genus. Nevertheless, the new species has eight to nine chambers of test, while only three-chamber forms are typical of the genus *Theocampe*.

Distribution. Upper Cretaceous, Upper Santonian–Lower Campanian, Perapedhi Formation, Cyprus.

Material. Seven specimens from type locality.

Subfamily Calocyclinae Haeckel, 1881

Genus Theocoronium Haeckel, 1887

Theocoronium subtriquetrus Bragina, sp. nov.

Plate III, figs. 2–4.

Etymology. Subtriquetrus Lat.—subtriangular.

Holotype. GIN 4840-5, Cyprus, Perapedhi Village, Upper Santonian–Lower Campanian, Perapedhi Formation

Description. The test with three chambers is of a helmetlike form. Its surface has numerous small spines. The cephalis is of the acropyramidal type. It is large and has small, sparsely distributed pores and a massive apical horn, longer than the cephalis. The vertical spine in the form of a small horn is two times shorter than the apical horn. The vertical horn is usually poorly preserved, because its basal part is hollow and thin-walled. In this case, the basal part of the vertical horn can be seen as a small hollow tube at the surface of cephalis. A well-developed incision is observed between the cephalis and thorax. The pores of the thorax are atypically circular, different in size and irregular in distribution. Spines D, L_1, and L_2 are distinct all over the cephalis surface. At the boundary between the thorax and abdomen they are long, extending outside the test. The proximal part of the abdomen is cylindrical. Well-preserved specimens have an abdomen with a narrow distal part, with nearly closed aperture. The wall of the abdomen has numerous pores, and sometimes is of a spongy texture. It covers the proximal parts of the spines like patagium.

Dimensions are as follows: Length of apical horn—90 μm; length of cephalis—50–60 μm; maximum width of cephalis—40–60 μm; maximum width of thorax—110–150 μm; length of test—250–350 μm, maximum width of test—120 μm (without spines).

Comparison. The new species differs from *Theocoronium ornatum* Bragina (Santonian–Lower Campanian of northwest Kamchatka) by the massive test with the more massive and long spines D and L, and by the absence of a collar at the distal part of abdomen (Bragina, 1991).

Distribution. Upper Cretaceous, Upper Santonian–Lower Campanian, Perapedhi Formation, Cyprus.

Material. Five specimens from type locality.

Family Stichocapsidae Haeckel, 1881

Subfamily Archaeodictyomitrinae Pessagno, 1976

Genus Stichomitra Cayeux, 1897

Stichomitra perapedhia Bragina, sp. nov.

Plate II, fig. 5, 6.

Stichomitra sp.: De Wever *et al.*, 1988, p. 171, Pl. 1, fig. 6.

Etymology. Perapedhi—village in central Cyprus.

Holotype. GIN 4840-4, Cyprus, Perapedhi Village, Upper Santonian–Lower Campanian, Perapedhi Formation.

Description. The test is strongly segmented, thick-walled, of wide conical shape, and consists of 4–5 chambers. The bulbous cephalis is large. The upper part of cephalis is smooth, bearing a rodlike apical horn equal in length to the cephalis. The lower part of the cephalis is moderately covered with irregularly arranged small circular pores. The summary length of the thorax and abdomen is equal to the length of the cephalis. Pores of the thorax, abdomen, and first postabdominal chamber are slightly bigger than pores of the cephalis. They have an unclear hexagonal arrangement with a weak tendency to form transversal rows. At the last chamber, pores are bigger and have a clear hexagonal arrangement. Pores are clustered in cells, and the surface of the test is moderately thorned.

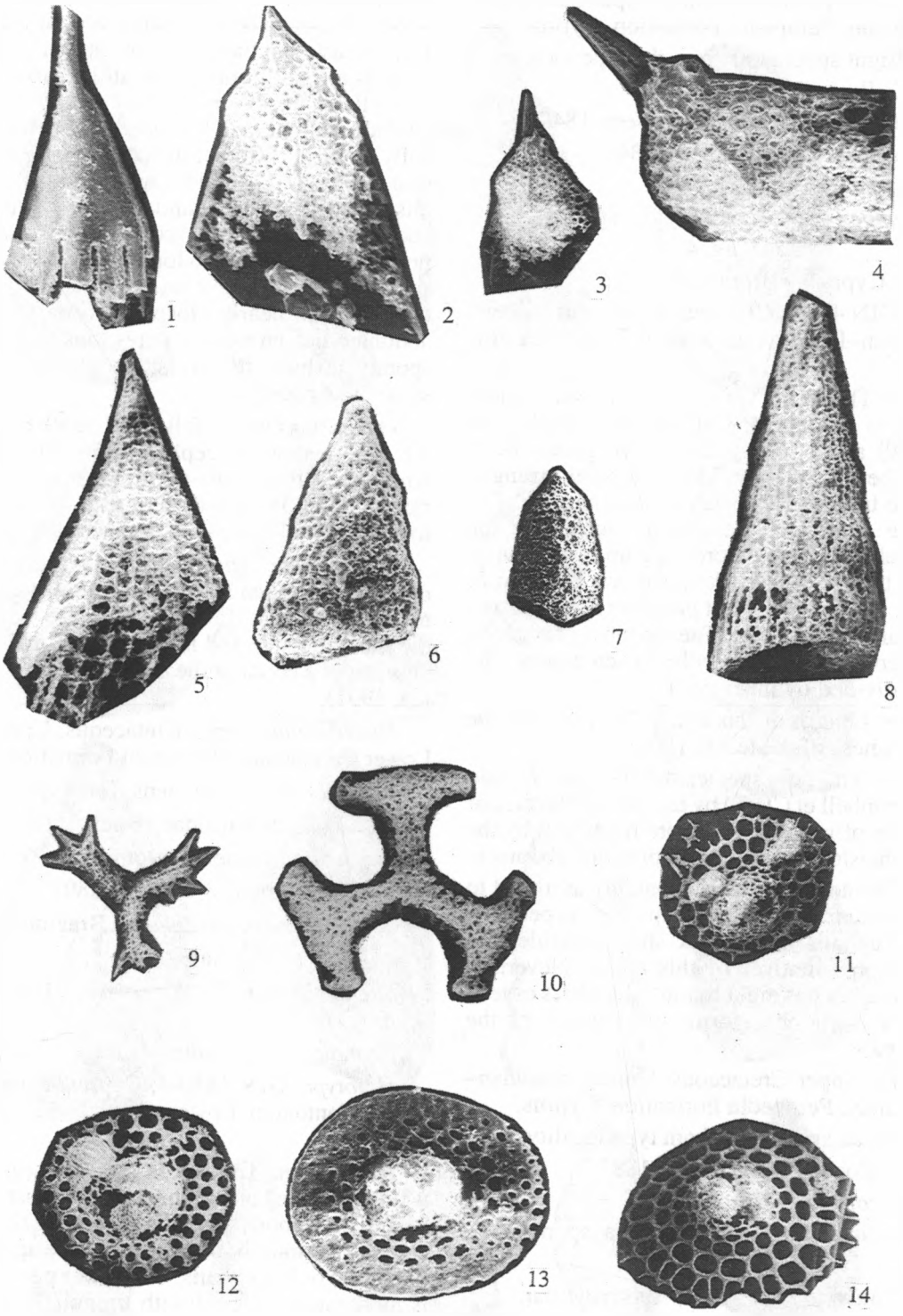


Plate III.

(1) *Afens lirioides* Riedel et Sanfilippo, ×150; (2–4) *Theocoronium subtriquetrus* Bragina, sp. nov.: (2, 4) ×200; (3) ×100. (5) *Bathropyramis* sp., ×150; (6) *Amphipyndax stocki* (Campbell et Clark), ×150; (7) *Stichomitra* sp. A, ×75; (8) *Dictyomitra densicostata* Pessagno, ×200; (9) *Hagiastridae* gen. indet., ×75; (10) *Halesium* sp. (Blome et Irwin, 1985), ×150; (11, 14) *Sciadiocapsa* sp., ×100; (12, 13) *Sciadiocapsa cansia* Foreman, ×100.

Dimensions are as follows: Length of apical horn—35 μm, length of cephalis—60 μm (with apical horn), maximum width of cephalis—50 μm, length of test—250 μm, maximum width of test—200 μm.

Comparison. The new species differs from *S. cechena* Foreman by the presence of apical horn, smaller length of the thorax and abdomen, thorned surface of test (due to pore clusters), wide open aperture, and by more massive and thick-walled test.

Distribution. Upper Cretaceous, Upper Santonian–Lower Campanian, Perapedhi Formation, Cyprus.

Material. Dozens of specimens from type locality.

Reviewers V.S. Vishnevskaya and I.A. Basov

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Paleeocene Reference Sections of Mangyshlak

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Abstract—The vertical distribution of macrofossils (echinoids, bryozoans, brachiopods, bivalves, and others), and planktonic and benthic foraminifers was studied in several Paleocene and Eocene sections of western Mangyshlak. The biostratigraphic subdivision of the sections and correlation with those of the Crimea and western Europe were carried out. Stratigraphy of the lower Paleocene (Danian and Montian), upper Paleocene (Selandian and Thanetian), lower Eocene (Ypresian), middle Eocene (Lutetian and Bartonian), upper Eocene (Priabonian), and lower Oligocene (Rupelian) was substantiated. The investigations were mostly devoted to the Paleocene.

Key words: *Mangyshlak, Paleocene sections, macrofossils, foraminifers, Danian, Montian, Selandian, Thanetian.*

In previous papers, we substantiated the validity of the subdivision of the Paleocene into a lower (Danian and Montian stages) and upper (Selandian and Thanetian stages) subseries (Naidin and Ben'yamovskii, 1988, 1989).

In the present work, the discussion on the Paleocene biostratigraphy is continued based on our field observations in western Mangyshlak, where Paleocene and younger deposits are widespread and well exposed. When describing the extent of the exposures in Mangyshlak, N.I. Andrusov wrote, "If in central Russia we should number the outcrops, here we would have to number the unexposed sites" (Andrusov, 1915, p. 111).

Danian deposits in Mangyshlak were revealed as early as the end of the nineteenth century (Semenov, 1899). Later on, Andrusov (1912, 1915) found that Danian bryozoan limestones in the southern Aktau are overlain with angular unconformity by a characteristic sequence of yellow sands with very large terebratulids. As the yellow sands overlie the Danian limestones and underlie nummulitic limestones (nummulites collected by Andrusov were determined by Duville), he assigned these terebratuloid sands to the lowermost Paleocene (Andrusov, 1915, p. 287).

The results obtained by Andrusov were of principal importance for understanding the Paleogene stratigraphy of Mangyshlak. Due to their occurrence mode and sharp lithologic distinction from Danian limestones, the terebratuloid sands are well studied in the southern Aktau. Vyalov (1945) called them Chagyrgan Sands. Liverovskaya later assigned them to the Sulukapy Formation (1960). A rich assemblage of bivalves (mainly oysters) and brachiopods allowed the Sulukapy Formation to be correlated with the Thanetian strata of the Crimea and Caucasus (Liverovskaya, 1960; Il'ina, 1963; Kovaleva, 1961).

Investigations carried out by G.S. Panteleev contributed much to the understanding of the Paleocene biostratigraphy of Mangyshlak. On the basis of the detailed study of sections in northern and southern Aktau, he revealed in the upper part of bryozoan limestones (earlier referred by all researchers to the Danian Stage) a molluscan complex characteristic of the Montian coarse limestones of Belgium. Panteleev described in detail the lower Paleocene bivalves from Mangyshlak and other Transcaspian regions (Tuarkyr, Krasnovodsk Peninsula, Bol'shoi Balkhan) in his monograph (Panteleev, 1974), which enhanced the reliability of his stratigraphic conclusions.

We should point out one more aspect of Panteleev's study which is important for Paleocene stratigraphy. Along with Montian mollusks, he collected echinoids among which M.M. Moskvin determined *Echinantus* typical of the Montian coarse limestones (Aliev *et al.*, 1967). Along with bivalves, echinoids are the most abundant macrofossils of the Upper Cretaceous and Paleogene in Mangyshlak. Hence, we seek to use macropaleontological remains of the Paleogene in Mangyshlak more completely. The macropaleontological characteristics of sections are based on determinations of the collected macrofossils. Bryozoans were identified by T.A. Favorskaya, brachiopods by M.V. Titova, crinoids by V.G. Klikushin, echinoids by M.M. Moskvin, corals by E.I. Kuzmicheva, serpulids by M. Egger (Germany), teeth of lamnoid sharks by V.I. Zhelezko, and large foraminifers (nummulitids) by A.A. Ashurov. V.A. Krasheninnikov consulted us on planktonic foraminifers of the Usak section. We express our deep gratitude to all of the researchers mentioned above. The published material of Panteleev (1974) on bivalves from the Kaurtakapy and Sulukapy sections were also used. Smaller benthic and plank-

tonic foraminifers were identified by V.N. Ben'yamovskii and L.F. Kopaevich.

The relationship between the macro- and micropaleontological characteristics of the Paleocene and Eocene startotypes of Mangyshlak was also an important problem. Some researchers earlier cited the lists of characteristic foraminifers from the Paleogene formations of Mangyshlak (Kuznetsova, 1958; Liverovskaya, 1960; Merklin *et al.*, 1960; Pleshcheev, 1970; Bykova, 1974; Pleshcheev *et al.*, 1981).

Some works present the description of characteristic species of smaller Danian benthic foraminifers and larger Ypresian–Lutetian foraminifers (Andrusov, 1915; Vasilenko, 1961; Il'ina, 1953; Barkhatova and Nemkov, 1964, 1965), while others deal with the Paleogene paleogeography of the Mangyshlak based on foraminifers (Kuznetsova, 1958; Nemkov, 1970).

Andrusov (1912, 1915) revealed the clear differences between the Paleogene deposits (beginning with the Paleocene terebratuloid sands identified by him) and underlying Danian bryozoan limestones. Later on, researchers long considered the Danian bryozoan limestones, uniform throughout the Mangyshlak mountains, as a single stratigraphic unit.

Paleogene deposits of Mangyshlak overlying these limestones show two peculiar features. First, they are represented by diverse facies, among which terrigenous varieties prevail. Therefore, much attention was given to the lithostratigraphic subdivision of the Paleogene deposits of Mangyshlak into individual formations. Second, the Paleogene deposits of the region are well characterized by macrofossils at all stratigraphic levels, in places, some of their groups (e.g., oysters) are rich in taxa. Unfortunately, the last feature was not fully considered. No biostratigraphic correlation with sections from other regions was carried out. Only incomplete data on the possible age of some stratigraphic levels of the Paleogene deposits of Mangyshlak were published. With the exception of Panteleev's book on Danian and Montian bivalves (1974), monographs on Paleogene macrofossils of Mangyshlak are unavailable. Only in a few works can one find a description and photographs of bivalves (Dylevskaya, 1913; Vyalov, 1945) and echinoids (Moskvin and Shimanskaya, 1981; Endelman, 1980a, 1980b) from the Paleocene deposits of Mangyshlak.

DESCRIPTION OF SECTIONS

Figure 1 shows locations of the studied sections. The Kaurtakapy, Sulukapy and Usak sections are graphically illustrated (Figs. 3–8). Sections are described from the bottom upward. Lists of macrofossils and microfauna are an essential part of the section characteristics. The species mentioned in the lists are cited in the text without the authors' names.

Kaurtakapy Section

The Kaurtakapy gorge crossing the North Aktau Ridge northeast of the Shetpe Settlement cuts a series of beds gently dipping northward. The beds are exposed more completely on the left slope of the gorge directly near the railway distance marker 293/294 km (Figs. 2–5).

Maastrichtian. Bed 1: chalk (fine-detrital limestone), white, slightly sandy; hard grounds are developed at the bed top. Organic remains characteristic of the upper Maastrichtian were collected from the adjacent Aksyirtau section (Naidin *et al.*, 1984); a change of macro- and microfauna at the boundary between the Maastrichtian and Danian (Naidin *et al.*, 1990a, 1990b) was recorded.

Danian. Bed 2: limestones, organogenic-detrital, light-gray, almost white, clear hard grounds are recorded 6.5 m above the bed base; bivalves *Pycnodonte (Phygraea) simile* (Pusch) and *Spondylus faxensis* Lundgren are associated with the benthic foraminifers *Osangularia lens, Stensioina whitei, Anomalina*

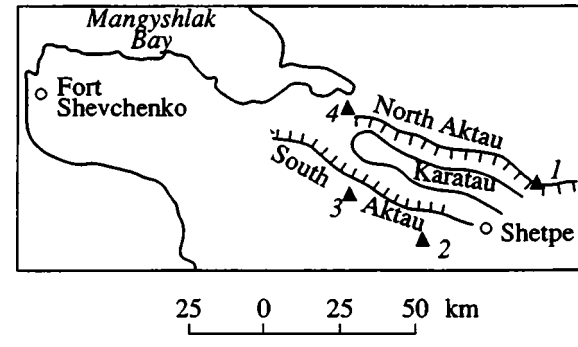

Fig. 1. Location scheme of the studied sections.
Sections: (*1*) Kaurtakapy; (*2*) Sulukapy; (*3*) Chat; (*4*) Usak.

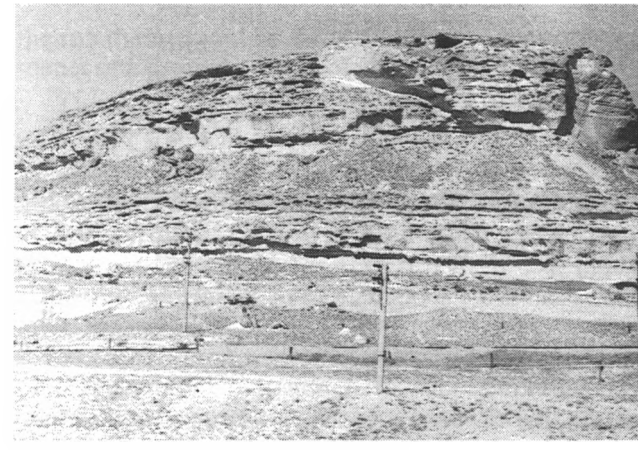

Fig. 2. General view of the Kaurtakapy Section.

danica, and *Heterolepa hemicompressus*. The thickness is 11–12 m.

Bed 3: limestones, organogenic-detrital, light gray, almost white, with hard grounds and black to dark-brown irregularly shaped cherty horizons. The sequence comprises the benthic foraminifers *Heterostomella gigantica, Alabamina wilcoxensis, Karreria fallax, Cibicidoides commatus*. At the top of the sequence, a triple horizon of cherts contains various macro- and microfauna groups. The upper part of the horizon comprises bivalves *P. (Ph.) simile, Liostrea nikitini* (Arkhangelskii), *Spodylus danicus* (Ravn), *Mediocardia taxensis* (Lundgren), and *Brachiodonte cottea* (Roemer); echinoids *Echinocorys obliqua* (Ravn), *E. sulcata* (Goldfuss), and *E. pentagonalis* Kongiel, rare brachiopods *Chatwinothyris* sp., and *Gyrosoria* sp.; and benthic foraminifers *Pyramidina crassa* and *Bulimina trigonalis* (=*Reusella paleocenica*). In the lower part of the horizon, the rare brachiopods *Ryorhynchia rionensis* (Antula) are present. The thickness is 63–64 m.

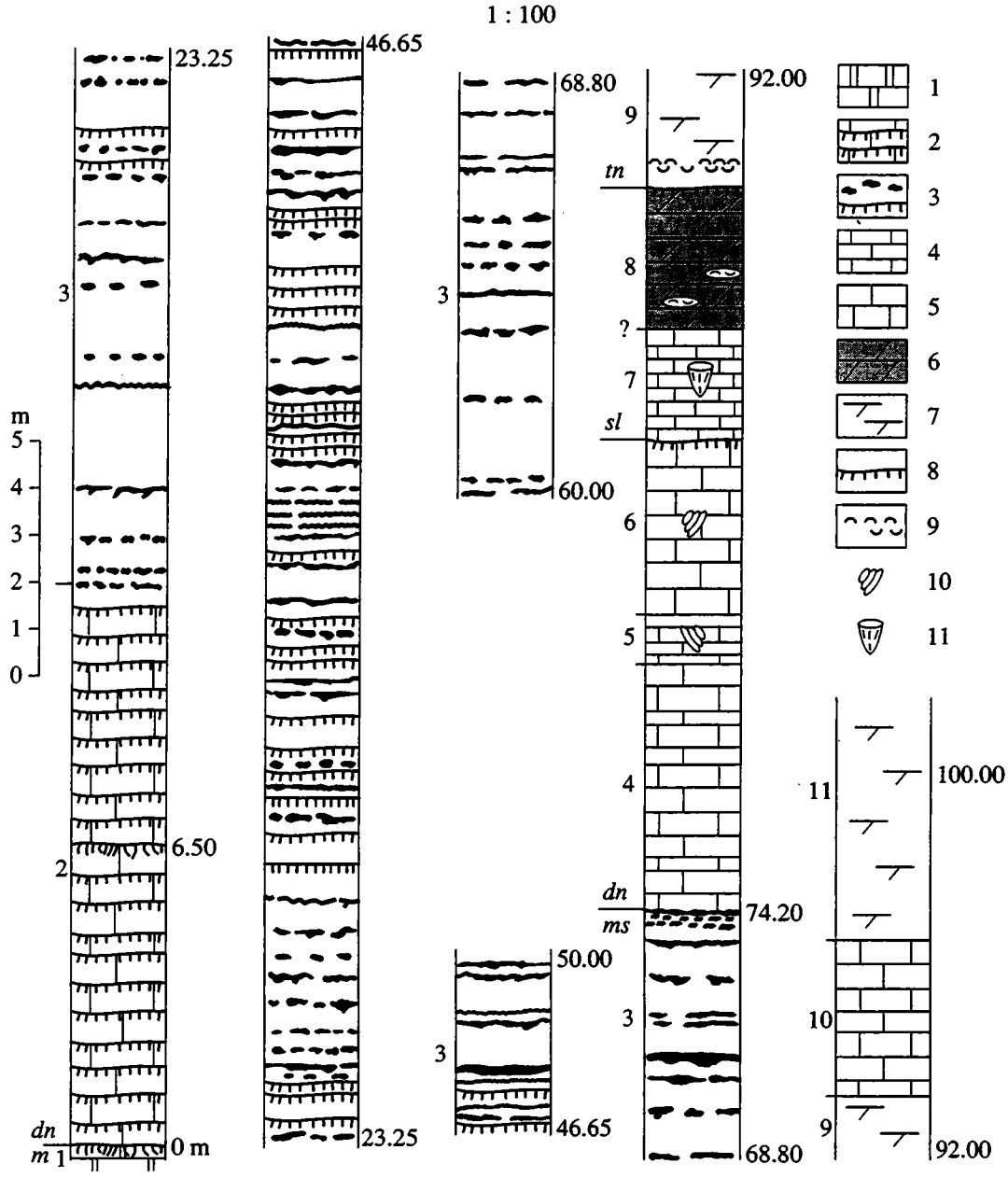

Fig. 3. The Kaurtakapy section. (1) chalk (small-detritus limestones); (2) organodetrital limestones with hard grounds; (3) organodetrital limestones with hard grounds and cherts; (4) different limestones; (5) slightly sandy limestones; (6) silty marls with brachiopods; (7) highly silty and sandy marls; (8) burrows of thalassinoids; (9) valves of oysters; (10) casts of large gastropods; (11) sponges. Numerals to the right of columns mark the height above the top of Bed 1. Abbreviations: (*m*) Maastrichtian, (*dn*) Danian, (*ms*) Montian, (*sl*) Selandian, (*tn*) Thanetian, (*yp*) Ypresian, (*lt*) Lutetian, (*br*) Bartonian, (*pr*) Priabonian, (*rp*) Rupelian.

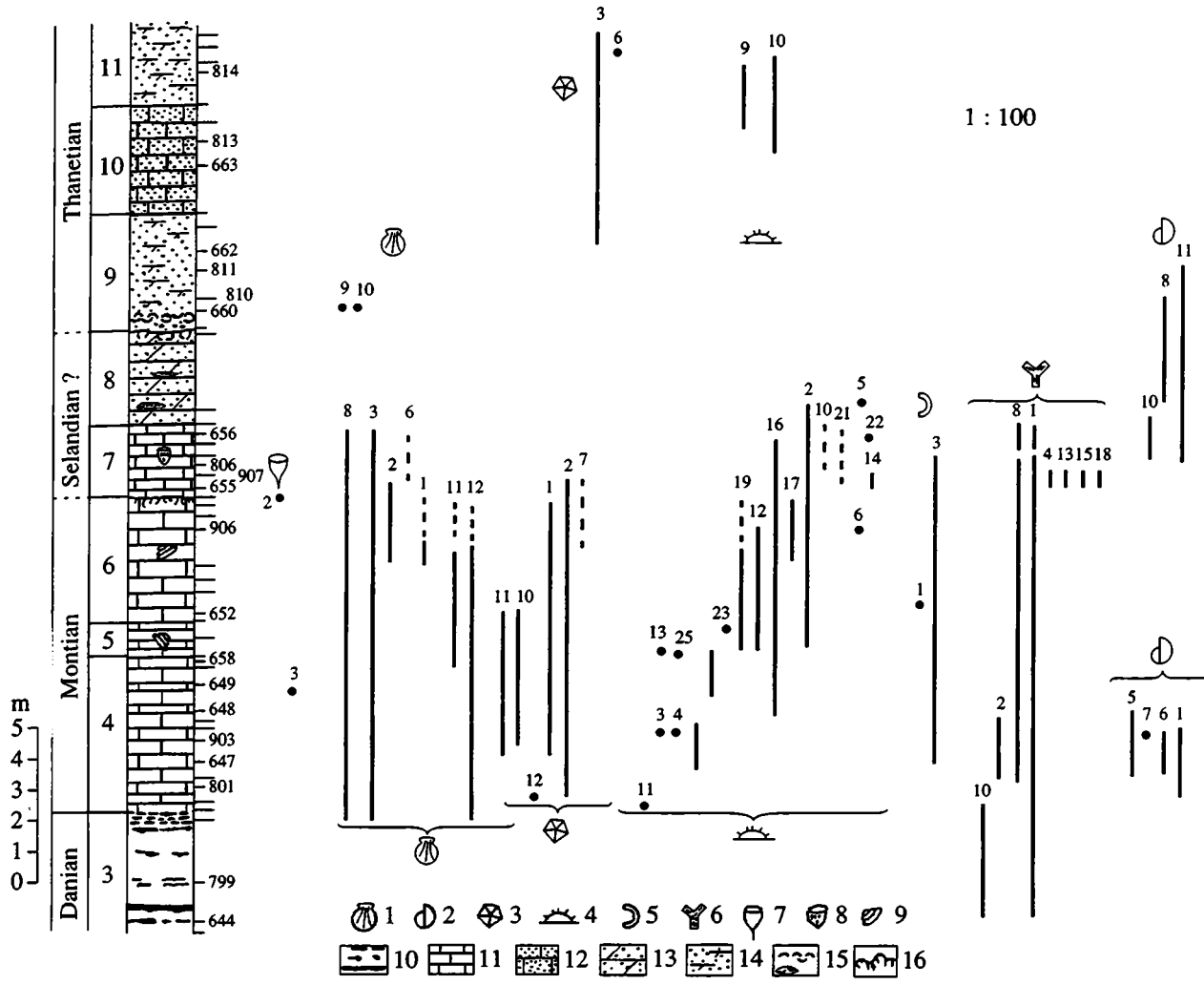

Fig. 4. The Kaurtakapy section. The Danian–Thanetian macrofauna.
(1) bivalves; (2) brachiopods; (3) crinoids; (4) echinoids; (5) serpulids; (6) bryozoans; (7) corals; (8) sponges; (9) casts of large gastropods; (10) organogenic-detrital limestones with cherts; (11) different limestones; (12) slightly sandy limestones; (13) highly silty and sandy marls; (14) silty marls; (15) oyster horizons (Bed 9), and lenses of large gastropods; and (16) hard grounds. For fossil remains shown in the figures see Appendix.

Montian. Bed 4: limestones, fine detrital, bryozoan, white and soft. On the right slope of the gorge, an intermittent interlayer with small black cherts is locally observed at the top of the bed. Echinoids *Echinanthus* aff. *issyanivensis*, Ech. aff. *pouechi*, *Neoglobator* aff. *houzeaui*, *Trematopygus* aff. *analis*, *Studeria crassa*, *S.* aff. *faberi*; benthic foraminifers *Cibicidoides lectus*, *Heterolepa bratus*, *Angulogerina wilcoxensis*; and planktonic foraminifers *Globoconusa koslowskii* are present. The thickness is 4.5–5 m.

Bed 5: limestones, fine detrital, bryozoan, platy, enclosing casts of large gastropods. The thickness is 1–1.2 m.

Bed 6: limestones, similar to those in Bed 4, porous in the upper part, with rare casts of large gastropods; at the top thalassinoid burrows filled with limestone fragments one present. The thickness is 3.8–4 m.

The top limestones of Bed 6 form a small structural terrace.

The assemblage of echinoids with *Neoglobator ovalis*, *N. houzeaui*, and *Echinocorys* aff. *ovliqua* is characteristic of Beds 5 and 6.

According to Panteleev (1974), Beds 4–6 contain remnants of bivalves *Acra montensis* Cossman, *Cucullaea montensis* Vincent, *Glycymeris duponti* Cossmann, *Venericardia excellens* Gorbach, *Lucina* (*Cavilucina*) *duponti* Cossmann, *L.* (*C.*) *lamellicardia* Cossmann, *Nemocardium ciplyense* Vincent, *Pitar duponti* Cossman, and *P. montensis* Cossman.

Beds 4–6 mark a step on the slope. They contain thalassinoid burrows.

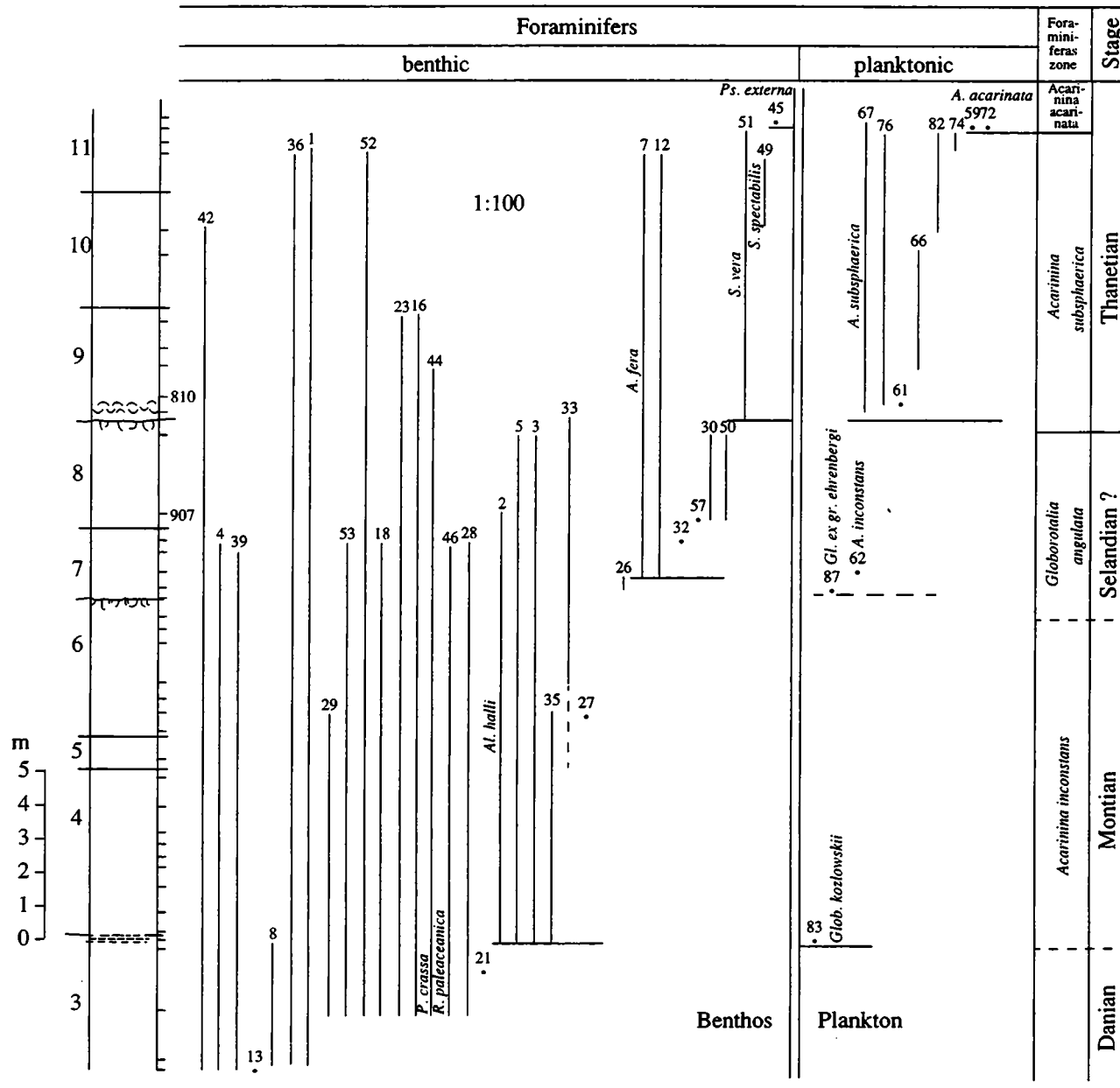

Fig. 5. The Kaurtakapy section. The Danian–Thanetian microfauna.

Selandian? Bed 7: limestones, grayish white, bluish gray on the weathered surface, containing silicified sponges at the base with rare fragments of small greenish-brown phosphorites. The third assemblage of echinoids *Pseudogibbaster tercensis, Neoglobator insolitus, Echinocorus cotteaui,* and *E. sulcata cognatus,* along with the benthic foraminifers *Valvulineria ravni, Gyroidina pontoni, Anomalina fera, Brotzenella acuta, Pullenia americana,* and planktonic foraminifers *Acarinina inconstans* and *Globorotalia* ex gr. *ehrenbergi* are present here. The thickness is 2.2 m.

Bed 8: marls, highly silty, light greenish gray, with lenses of large brachiopods; crayfish burrows and scattered small fragments of light-gray phophorites and phosphorized sponges occur in the upper part of the bed; marls are exposed at the steep part of the slope. The thickness is 3 m.

Thanetian. Bed 9: marls, highly sandy, light greenish gray; their thick layers form benches on a gentle slope; small scattered fragments of black rocks (phosphorites?) occur in the lower part of the bed; a layer filled with oysters, rare brachiopods, and still rarer spondyloids occur at the base of the bed. The thickness is 3–3.5 m.

It seems that *Arca morrisi* (Sowarby), assigned by Panteleev (1974) to the Kaurtakapy section, was described from this bed.

Bed 10: limestones, slightly sandy with bright yellow small spots; they are almost white when fresh, and grayish-yellow when weathered; the interval is marked by small niches and benches. The thickness is 3–3.5 m.

Beds 9 and 10 contain the benthic foraminifers *Spiroplectammina vera, S. spectabilis* and planktonic foraminifers *Globigerina nana, G. velascoensis, G. incisa, Acarinina subsphaerica, A. soldadoensis*, and *A. djanensis*.

Bed 11: marls, silty, dusty, yellowish gray; they form a gentle part of the slope and contain the benthic foraminifers *Pseudogaudryina externa* and planktonic foraminifers *Globigerina crassa*, and *Acarinina acarinata*.

The forth assemblage of echinoids *Echinocorys schwetzovi* and *E.* cf. *schwetzovi* is associated with Beds 10 and 11.

Sulukapy Section

The section begins at the mouth of the Sulukapy Gorge, 1.5 km from the Tushebek Settlement. According to Andrusov (1915), the gorge represents a narrow canyon. It cuts the South Aktau Ridge, where the first interval of the section consists of the following Maastrichtian, Danian, and Montian beds (Fig. 6):

Maastrichtian. Bed 1: limestones, organogenic-detrital, light yellowish gray, with hard grounds; hard sandy limestones with *Crania* valves and the last rostra of belemnites occur in the upper part of the bed; hard limestones form a small bench.

Danian. Bed 2: limestones, organogenic-detrital, light yellow, or pink and pale yellow, with hard grounds; small phosphorite nodules occur at the base. The bed contains rare, well-preserved macrofossil remains of *Pyctodonte* (*Phygraea*) *similis* (Pusch) and other oysters, along with teeth of *Jackelotodus spaeri* Dartevelle et Casier and *Odontaspis* sp., which are common in the upper part of limestones. The thickness is 40–42 m.

Montian. Bed 3: limestones, organogenic-detrital, light pinkish gray, with hard grounds, scattered gravel and small phosphorite pebbles often concentrated in lenticular bands, and layers with bivalves (mainly small oysters). Teeth of sharks *Striatolamia* aff. *striata* Winkler, *Chararodu* sp., and *Odontaspis* sp. are present in limestones, which are arenaceous in the lower part and obliquely laminated in the upper part. A horizon of large nodule up to 20–25 cm in size and with a porous surface and brown phosphorite concretions occurs at the top of the bed (it is probably the destructed hard ground). The thickness is 13–14 m.

Monsian(?). Bed 4: limestones, organogenic-detrital, similar to those in Bed 3, and obliquely laminated in the upper part hard grounds are rare. The thickness is 6 m.

In Beds 3 and 4, Panteleev (1974) defined the bivalves *Lucina* (*Cavilucina*) *lamellicardia* (Cossmann), *Pitar duponti* Cossmann, and *P. montensis* (Cossmann).

The detritus fraction of Beds 2–4 is represented by bryozoan remains. The honeycomb weathering patterns are developed on the surface of limestones of Beds 1–4 and exposed on the gorge walls, especially on the left one. A continuous sequence of Beds 1–4 not concealed by honeycomb weathering is observed on the right wall of the gorge; at the sequence base, the remnants of an ancient aryk (irrigation ditch) occur in limestones.

The Sulukapy Gorge, which cuts the hard limestones of Beds 1–4, widens southward and passes into the wider Sartagan Valley. The second interval of the section exposed in this valley is composed of the Thanetian and higher layers (Fig. 7):

Thanetian. Bed 5: sandstones, calcareous, arenaceous, more compact at the base. Throughout the bed, they are yellowish gray, with scattered rounded phosphorite nodules of 3 cm in size and the oysters *Pycnodonte* (*Phygraea*) *sullukapensis* (Vialov), *P.* (*Ph.*) *nomada* (Vialov), and *P. sinzovi* (Netschaew). Bryozoans are abundant here, while brachiopods *Terebratula bisinuata* Lamarck are also numerous, and *Neoliothyrina fallax* (Lundgren) are more rare and found only at the base of the bed.

Several layers of large oysters occur 3.5–4.1 m above the bed base, and some contain burrows of *Cliona* and fragments of phosphoritic rocks bored by pholads; at the top of the bed, phosphorite pebbles are present (some of them are bored by pholads). The thickness is 32–33 m.

Bed 5 is observed and described on the right slope of the Sartagan Valley near the Sulukapy Gorge outlet.

Ypresian. Bed 6: limestones, white, with large nummulites (*Nummulites distans*), very hard. The abundant hard grounds are responsible for the porous surface of limestones; some of the hard grounds are pink and rusty yellow. The thickness is 4.5–6 m.

Lutetian?. Bed 7: limestones, nummulitic, light gray to almost white, hard, with large echinoids and foraminifers *Nummulites murchisoni, N. irregularis, N. atacicus, N. anomalis*, and *N. rotulatus*. The thickness is 4–5 m.

Lutetian. Bed 8: sands and loose sandstones with large thick-shelled *Gryphaea rarilamella* and nummulites similar to those from the previous bed. In the upper part of the bed, planktonic foraminifers *Globigerina boweri*, and *G. ratusa* occur. The thickness is 7 m.

Bartonian. Bed 9: limestones and marls, light gray to almost white. Planktonic foraminifers *Globigerapsis subconglobatus*, and *Globigerina subtriloculinoides* are present. The thickness is 1.5–1.7 m.

Bed 10: sandstones, highly calcareous, light and dark brown or greenish gray in the upper part, dense, massive. The thickness is 6 m.

Priabonian. Bed 11: marls, sandy, white-pink, compact; the rocks contain the benthic foraminifers *Plan-*

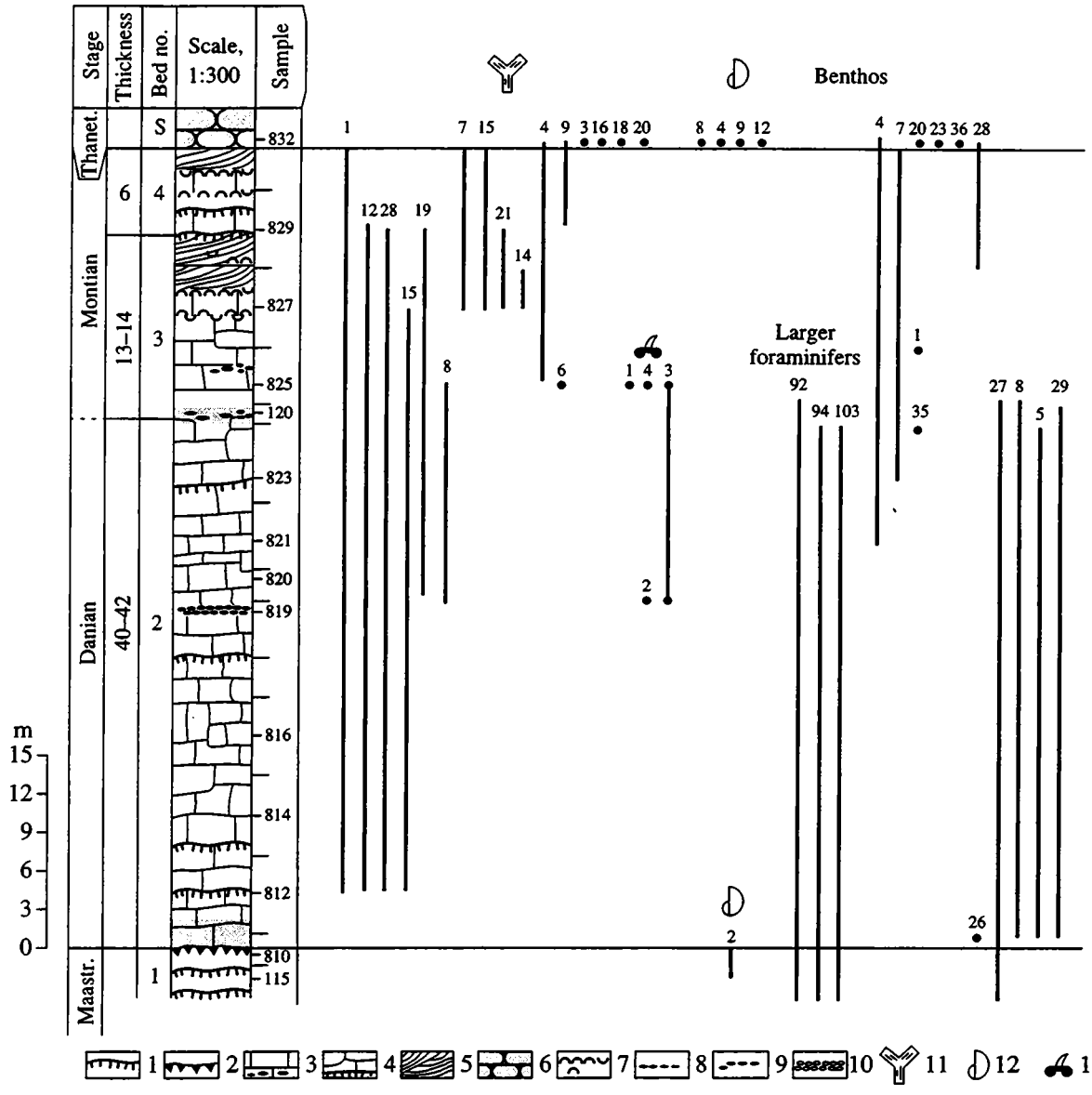

Fig. 6. The Sulukapy section. The Danian–Montian.

(1) light yellowish-gray organodetrital limestones with hard grounds; (2) hard sandy limestones at the top of Bed 1; (3) sandy limestones with small phosphorite nodules in the lower part; (4) light-yellow, pinkish, and pale-yellow, organodetrital limestones with hard grounds; (5) obliquely laminated organodetrital limestones; (6) calcareous limestones; (7) small bivalves (mainly oysters), single, and forming lenses and inconsistent layers; (8) sandy limestones with phosphorites; (9) segregations and single small phosphorites; (10) interlayer with contiguous hard grounds, small phosphorite and limestone pebbles; (11) bryozoans; (12) brachiopods; and (13) teeth of selachians.

ulina costata, and *Uvigerina jacksonensis* along with planktonic foraminifers *Globigerapsis tropicalis.* The thickness is 3 m.

Bed 12. marls, clayey in the upper part, grayish white with a greenish tint, bearing the benthic foraminifers *Brotzenella taurica* and *Bolivina antegressa.* The thickness is 8 m.

Rupelian. Bed 13: clays, gray to greenish gray, in places with a lilac or violet tint, scaly-laminated, enclosing the benthic foraminifers *Spiroplectammina carinata oligocenica, Gyroidina soldanii, Cibicidoides oligocenicus, C. expertus,* and *Bolivina missisippiensis.* The thickness is 11 m.

Beds of the Sulukapy–Sartagan Section show monoclinal southward dipping with a dip angle of 10°–12° in beds 1–4: limestones of Bed 4 are truncated by gently dipping Bed 5 and overlying strata. Upsection, the dip angle decreases down to 2°–4°, and the thickness of Bed 5 slightly increases.

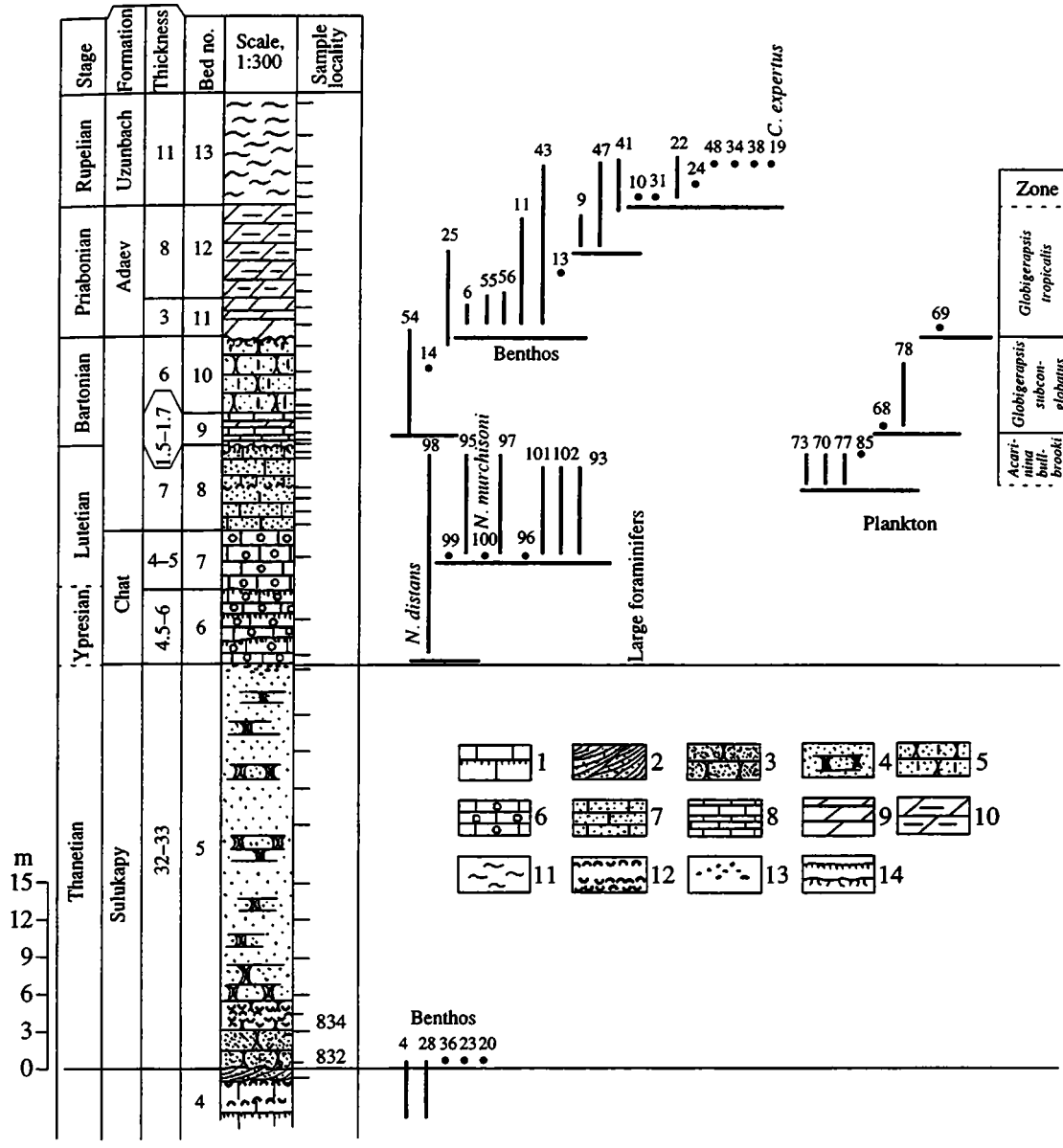

Fig. 7. The Sulukapy section. The Thanetian–Rupelian.
(1) organodetrital limestones; (2) obliquely laminated organodetrital limestones; (3) yellowish-gray sandstones (the base of Bed 5); (4) loose sandstones; (5) calcareous sandstones; (6) nummulitic limestones; (7) sandy limestone; (8) chalk-like limestones and chalk; (9) marls; (10) clayey marls; (11) clays; (12) accumulations of bivalve remains, mainly of oysters; (13) phosphorites; and (14) hard grounds.

Limestones of Bed 4 have a scarped surface. The relationship between Danian–Montian limestones (Beds 2–4) and Thanetian sandstones (Bed 5) is similar to that between Danian–Montian limestones and Thanetian–Ypresian marls and clays in the southwestern Crimea Mountains.

Chat Section

The Chat section is situated near the Chat Well, 7 km south of the Tauchik Settlement.

Maastrichtian. Bed 1: chalky white.

Danian–Montian. Bed 2: limestones, organogenic-detrital with hard grounds. The lower 12–13 m of limestones contain no cherts; overlying limestones (25–27 m) comprise irregular concretions of cherts, the largest of which ("boulders" 0.6–0.7 m across) occur in the upper part of this interval. The upper part of the bed is also free of cherts (the Monsian, 20–24 m). Platy limestones are found in some parts of the section, and at the top of Bed 2. Benthic foraminifers *Stensioeina lens* and *Anomalina danica* are present; *Heterolepa bratus* occurs in the uppermost part of the bed.

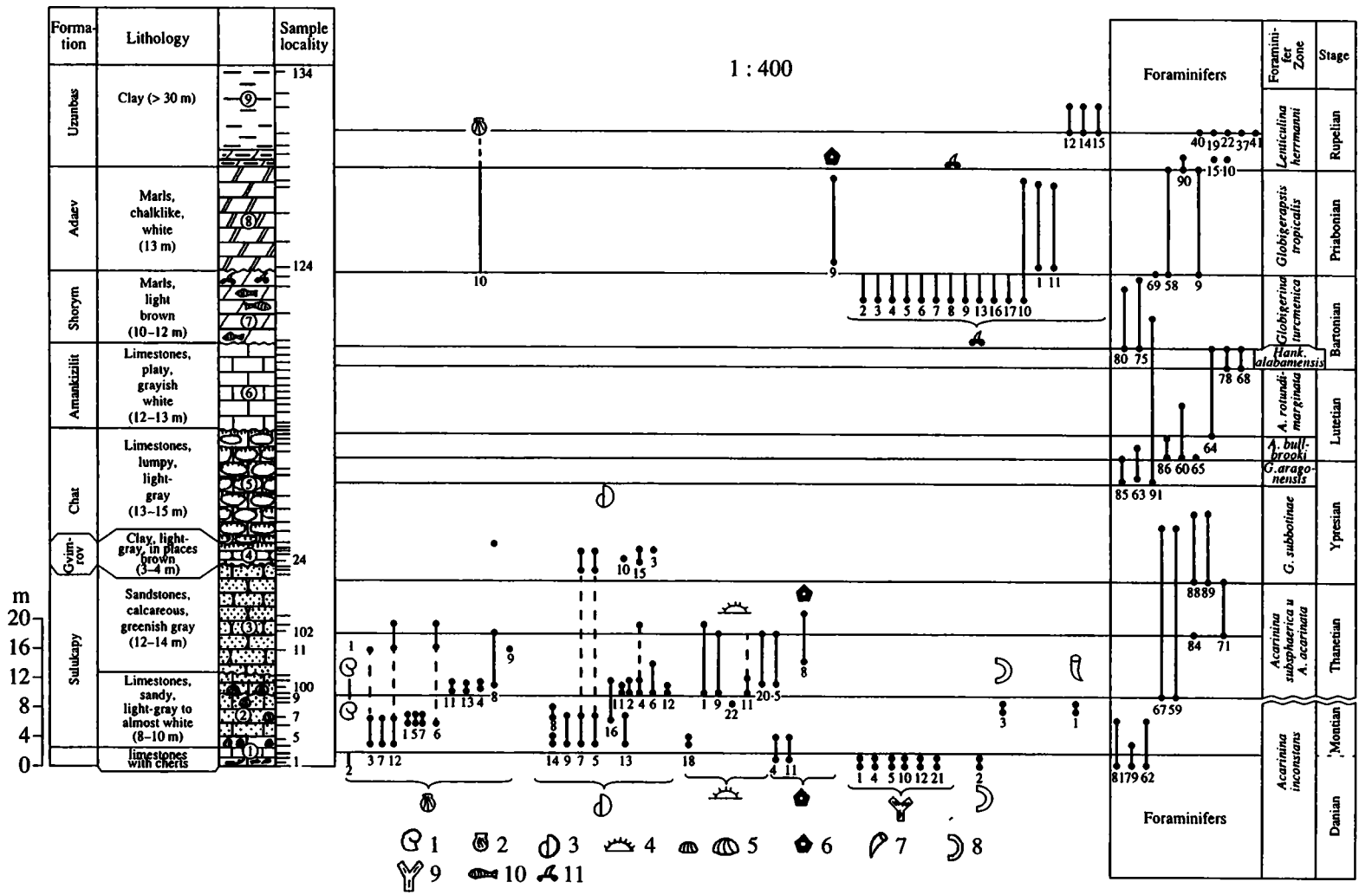

Fig. 8. Usak section. The Montian–Rupelian. (1) nautilids; (2) bivalves; (3) brachiopods; (4–5) small and large echinoids: (4) in the lithological column, (5) in the lithological column; (6) crinoids; (7) corals; (8) serpulids; (9) bryozoans; (10) fish remains (in the lithological column); and (11) teeth of selachians. Abbreviations: *A.*—*Acarinina*, *G.*—*Globorotalia*, *Hantk.*—*Hantkenina*.

Thanetian? Bed 3: sands, rusty green, with fragments of phosphorites, phosphorized rocks, and marks of stone borers. The thickness is 0.2–0.4 m.

Thanetian. Bed 4: sands, compact, greenish gray, exposed in the vertical walls of outcrops. Benthic foraminifers *Valvulineria ravni, Eponides lunatus, Anomalina fera, Cibicides reinholdi*, as well as planktonic foraminifers *Globigerina nana, G. velascoensis, Acarinina subsphaerica,* and *A. acarinata* are present. The thickness is 5–5.5 m.

Ypresian? Bed 5: sands, calcareous, light greenish gray, laminated, dusty, with pockets and fissures at the top filled with phosphorites of the overlying bed. The large foraminifers *Discocyclina* ex gr. *frabaensis* are observed. The thickness is 4 m.

Ypresian. Bed 6: limestones, nummulitic, yellowish white, with hard grounds, porous. The rocks contain dark-brown phosphorite in the lower part and foraminifers *Nummulites anomalis, N. bolcensis* Munier-Chalmas, *N. murchisoni,* and *N. irregularis*. The thickness is 6 m.

Ypresian? Bed 7: limestones, slightly sandy, light gray to almost white, with the rare planktonic foraminifers *Acarinina marksi*.

Usak Section

The Usak section is situated west of the Aktau-Buzachi highway, near the Usak wells (Fig. 8).

The lower larger part of the calcareous limestones of Bed 2 contain very rare and poorly preserved *Echinantus* remains. In the upper part of this interval, rare and poorly preserved casts of *Pseudogibbaster* cf. *tercensis* are found. The mentioned echinoids are not shown in Fig. 8. Based on *Echinanthus* remains, this part of Bed 2 is attributed to the upper part of the Montian, as it contains elements of the *Micraster tercensis* assemblage. In this interval, planktonic foraminifers *Acarinina subsphaerica* and *A. acarinata* were also found.

North of the wells, in the rest of the Chat valley, deposits between the Danian and Thanetian intervals, are characterized by a more diverse echinoid assemblage (Endelman, 1980b).

SUBSTANTIATION OF BIOSTRATIGRAPHIC SUBDIVISION

The data presented allow us to resume and verify the biostratigraphic correlation of Mangyshlak and western European sections commenced by Panteleev and Moskvin.

We believe that in the Paleocene and early Eocene, as in the Late Cretaceous, the Mangyshlak was a part of the European paleobiogeographical realm. From this viewpoint, the correlation of Mangyshlak and western European sections, including Paleogene stratotypes is easier.

The upper part of the white bryozoan limestones attributed by Semenov (1899) and Andrusov (1915) to the Danian was assigned by Panteleev (1974, Figs. 10 and 11) and Aliev *et al.* (1967, Plates 1 and 2) to the *Echinanthus* Beds of the Montian Stage s. str. They also singled out transitional beds correlatable with the Tuffeau Ciply of Belgium, which terminate the Maastrichtian of Mangyshlak. At present, Belgian researchers assign the Tuffeau Ciply to the Danian (Naidin and Ben'yamovskii, 1988, Figs. 3, 5, and 6).

In the northern Aktau, the base of the Montian was distinguished by the disappearance of cherts. In the Kaurtakapy section (Figs. 3 and 4), it corresponds to the base of Bed 4. In the southern Aktau (Sulukapy section), Beds 3 and 4 are attributed to the Montian. It seems that the base of the Montian should be referred to a higher level. T.A. Favorskaya reasoned that the base of Bed 4 (Sample 801) of the Kaurtakapy section corresponds to the middle part of Bed 3 (Sample 827) of the Sulukapy section. At this level, a mass appearance of *Coscinopleura angusta minor* Vogt was recorded.

Beds 4–6 of the Kaurtakapy Section contain two assemblages of echinoids. The presence of Danian echinoids *Studeria* and *Echinocorys* aff. *obliqua* was observed. Panteleev singled out the *Echinanthus* Beds of the Montian Stage s. str. as corresponding to Beds 4–6. Bed 7 terminates the echinoid assemblage, and Moskvin correlates it with the *Micraster tercensis* assemblage defined by Villatte (1962) in the Lesser Pyrenees. Moskvin (1983) assigned *Micraster tercensis* Cotteau along with some other forms to a new genus, *Pseudogibbaster* Moskvin.

Above the Danian and under the Thanetian limestones of the Lesser Pyrenees, Villatte (1962, pp. 42, 43) distinguished a member consisting of five beds of limestones, marls, sandstones and sands. Four lower beds comprise *Echinanthus*, whereas the fifth upper bed contains the *Micraster tercensis* assemblage. Stratigraphically, the member should be correlative to the Montian strata of Belgium. However, a direct correlation is complicated, because the mollusks and corals in the Montian strata of the Lesser Pyrenees are characteristic of the Thanetian beds in the Paris basin. Villatte (1962, p. 294) cited for Beds 1–5 the planktonic foraminifers *Globigerina triloculinoides, Globorotalia angulata, G. pseudobulloides, G. longata*, and *G.* cf. *acuta*. In her opinion, this assemblage is characteristic of the lower part of the *Globorotalia angulata* Zone of Loeblich and Tappan's scheme (Loeblich and Tappan, 1957). Panteleev (1970) pointed to problems, which appear when correlating the Montian of the Lesser Pyrenees with the Montian of Belgium. We consider Beds 1–4 of the Lesser Pyrenees as corresponding to the Montian in *Echinanthus* forms and micropaleontological data.

In the Usak section, Bed 1 and the lower part of Bed 2, where the planktonic foraminifers *Acarinina*

inconstans are present, correspond to the Montian. There are sufficient grounds to believe that the *Acarinina inconstans* Zone spans the uppermost horizons of the Danian and Montian (Naidin and Ben'yamovskii, 1988, 1989).

In the context of the current Paleocene planktonic foraminifer zonation, the association described by Villatte appears to be of mixed composition (Shutskaya, 1970; Krasheninnikov, 1982; Berggren *et al.*, 1985). It comprises *Globigerina triloculinoides* and *Globorotalia pseudobulloides*, which occur in the lower Paleocene (Danian–Montian), and *Globorotalia angulata, G. aequa,* and *G. acuta*, which are widespread in the upper Paleocene–lowermost Eocene and not encountered in the lower Paleocene. Bed 5, with the *Micraster tercensis* assemblage found in the Lesser Pyrenees, should be assigned to the Selandian.

In the Kaurtakapy section, Bed 7 with the third echinoid assemblage bearing *Pseudogibbaster tercensis* also corresponds, in our opinion, to the Selandian. On the basis of foraminifers of the *Globorotalia angulata* Zone, this bed is correlative with Bed 5 of the Lesser Pyrenees. Bed 7 comprises an association of the benthic foraminifers *Gyroidina pontoni, Valvulineria ravni,* and *Pullenia americana.* These forms are described by Brotzen (1948) from the Selandian deposits of Sweden. Foraminifers *Globorotalia* ex gr. *ehrenbergi* are encountered in this bed. The species *Globorotalia ehrenbergi* is characteristic of the *Globorotalia angulata* Zone, which is assigned to the Selandian (Krasheninnikov, 1982; Naidin and Ben'yamovskii, 1989; Berggren *et al.*, 1985).

Bryozoans *Puncturiella* (*Puncturiellina*) *cava* and brachiopods *Terebratula bisinuata*, which are characteristic of higher horizons of the Thanetian, appear in Bed 7.

The *Echinocorys cotteaui* species, typical of the *Micraster tercensis* assemblage, was revealed in the lower part of Bed 8 of the Kaurtakapy section. We refer the bed to the Selandian.

Paleogene sections in the southwestern Crimea Mountains, where the west European stages were reliably distinguished, are very important for correlation (Naidin and Ben'yamovskii, 1994).

The similarity of the Danian–Montian assemblages from the Crimea and Mangyshlak is beyond question (Gorbach, 1972; Moskvin, 1964; Moskvin and Naidin, 1964; Poslavskaya and Moskvin, 1964).

In Mangyshlak, as in the Crimea, overlying deposits rest with angular unconformity on the Danian–Montian sequence, where the hiatus corresponding to the largest part of the Selandian is observed (Naidin and Ben'yamovskii, 1989, Fig. 5).

The Thanetian bivalve assemblages from both regions are similar. They include *Pychnodonte* (Phygraea) *antiqua* (Schwetzow), *Liostrea reussi* (Netschaew), *Chlamys prestwichi* (Morris), and *Arctica morissi* (Sowerby).

For the Thanetian, *Pychnodonte* (*Phygraea*) *antiqua* is the archistratigraphic fossil important for distant correlations within the European paleobiogeograpgic realm. This oyster is found not only in the Crimea and Mangyshlak, but also in the Tajik depression, eastern Turkmenia, Abkhazia region, northern Ukraine, Bulgaria and Poland (Naidin and Ben'yamovskii, 1989, p. 26).

Beds 9–11 of the Kaurtakapy section, as well as the upper part of Bed 2 and Bed 3 of the Usak section, and Bed 4 of the Chat section contain planktonic foraminifers *Acarinina subsphaerica, A. acarinata, Globorotalia aequa, Globigerina nana,* and *G. velascoensis*, which are typical of the Thanetian sediments in the Crimea, Caspian region, Ciscaucasia, and Tajik depression.

Ypersian bivalves of the Crimea and Mangyshlak seem to be less similar, which is probably a consequence of insufficient studies in Mangyshlak.

Gryphaea rarilamella described by Dylevskaya from the Sulukapy section in 1913 is a common fossil of the Ypresian deposits in the Crimea and Mangyshlak. This giant oyster is also recorded in Ypresian deposits in northern France and western Belgium.

Vyalov (1986) inferred that Paleogene oysters are important for the subdivision of sections and distant stratigraphic correlations which was proven by other researchers (Dmitriev, 1965; Naidin and Ben'yamovskii, 1989).

The common character of nummulitic associations of the Crimea and Mangyshlak and their similarity to nummulites of the European paleobiogeographic realm were discussed by Nemkov (1970), as well as by Barkhatova and Nemkov (1964, 1965). Our data confirmed their inferrences.

CONCLUSION

1. The vertical distribution of macrofauna (echinoids, crinoids, bryozoans, brachiopods, bivalves, and others), planktonic and benthic foraminifers was studied in the Kaurtakapy (North Aktau Range), Sulukapy, and Chat (South Aktau Range) sections, as well as in the Usak section on the Kochak Bay coast.

2. Based on these studies, the biostratigraphic subdivision of the sections was substantiated and the deposits were discriminated into the Lower Paleocene (Danian and Montian), Upper Paleocene (Selandian and Thanetian), Lower Eocene (Ypresian), Middle Eocene (Lutetian and Bartonian), Upper Eocene (Priabonian), and Lower Oligocene (Rupelian) units.

3. Paleocene deposits consist of two stratigraphic units separated by a small, but consistent angular unconformity; Andrusov was the first to establish this fact (1915). The lower unit is represented by the Danian and Montian limestones, and the upper one comprises the Selandian, Thanetian, Ypresian and Lutetian–Rupelian deposits. Lithologically, this is a very complicated fossiliferous sequence, which was subdivided earlier

into many formations on a lithostratigraphic, and not biostratigraphic basis. A stratigraphic hiatus, the most part of which seems to span the Selandian, is marked by angular unconformity.

4. The use of macro- and micropaleontological data allowed us to correlate the Mangyshlak sections with those in the Crimea and western Europe. The stratigraphic importance of some bivalves (oysters, in particular) and echinoids was confirmed.

5. It is essential that the Paleogene is subdivided into stages on the basis of the stage stratotypes of western Europe (Naidin and Ben'yamovskii, 1988, 1989). As before, we believe that subdivision into stages, and not into zones is a matter of first priority. A similar inference was introduced by Morozova (1959, p. 289), an eminent Russian micropaleontologist.

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APPENDIX

Macrofauna

Nautilids

Usak section (Fig. 8) 1—*Atiroidea nephela* (Korobkov).

Bivalves

Kaurtakapy section (Fig. 4) (1) *Crassatella* cf. *faxensis* Ravn; (2) *Gryphaeostrea lateralis* (Nilsson); (3) *Gr.* cf. *lateralis* (Nilsson); (4) *Lima* cf. *holzapfeli* (Mantell); (5) *Lima* sp.; (6) *Liostrea* cf. *reussi* Netschaew; (7) *Plicatula* sp.; (8) *Pyctodonte* (Phygraea) *simile* (Pusch); (9) *P.* (*Ph.*) *antiqua* (Schwetzow); (10) *P.* (*Ph.*) *frauscheri* (Traub); (11) *Spondylus danicus* Ravn; (12) *Sp. dutempleanus* d'Orbigny; (13) *Teredo* sp.

Usak section (Fig. 8). (1) *Chlamys prestwichi* (Morris); (2) *Chl.* sp. *undulata* (Nilsson); (3) *Labrostrea labrum* (Vialov); (4) *L.* cf. *labrum* (Vialov); (5) *Laevicardium* sp.; (6) *Liostrea reussi* (Netschaew); (7) *Meiocardia* sp.; (8) *Pycnodonte* (*Phygraea*) *antiqua antiqua* (Schwetzow); (9) *P.* (*Ph.*) *antiqua sartaganica* (Vialov); (10) *P.* (*Ph.*) *brogniarti questeleti* Nyst; (11) *P.* (*Ph.*) *sinzowi* (*Netschaew*); (12) *Spondylus* sp.; (13) *Teredo* cf. modica Deshayes.

Brachiopods

Kartakapy section (Fig. 4)

Sulukapy section (Fig. 6). (1) *Gryhus* aff. *bosoktiensis* Nechrikova; (2) *Isocrania* cf. *poselti* Rosenkrantz; (3) *Neoliothyrina fallax* (Lundgren); (4) *N.* cf. *mobergi* (Lundgren); (5) *Neoliothyrina* sp.; (6) *Riorhynania rionensis* (Anthula); (7) *Riorhynchia* sp.; (8) *Terebratula bisinuata* Lamarck; (9) *T.* aff. *bisinuata* Lamarck; (10) *T.* cf. *bisinuata* Lamarck; (11) *Terebratulina* aff. *striata* Wahlenberg; (12) *Terebratulina* sp.

Usak section (Fig. 8). (1) *Basiliola* cf. *incurva* (Schlotheim); (2) *Eolocasella mangyshlakensis* Iljina; (3) *Erymnaria polymorpha* Cooper; (4) *Gryphus bajanianus* Davidson; (5) *G. besoktiensis* Nechrikova; (6) *G.* ex gr. *hilarionis* (Menegh.); (7) *G. hilarionis* (Menegh.); (8) *Orientothyris* cf. *elegantis* (Nechrikova); (9) *O. elegantis* (Nechrikova); (10) *Streptaris* aff. *streptimorpha* Cooper; (11) *Terebratula bisinuata* Lamarck; (12) *T.* cf. *magna* Nechrikova; (13) *T.* ex gr. *fallax* Lundgren; (14) *Terebratulina aksyrtaensis* Nechrikova; (15) *T. emarginata* Risso; (16) *T. tauremarginata* Sacco.

Echinoids

Kaurtakapy section (Fig. 4)

Usak section (Fig. 8) (1) *Brissopneustes* sp. nov.; (2) *Cyclaser* sp.; (3) *Echinantus* aff. *issyanivensis* (Klein); (4) *Ech.* aff. *pouechi* Cotteau; (5) *Echinocorys cotteaui* Lambert; (6) *E.* aff. *obliqua* Ravn.; (7) *E. schwetzovi* Moskvin et Shimanskaja; (8) *E.* cf. *schwetzovi* Moskvin et Shimanskaja; (9) *E. sulcata* Goldfuss; (10) *E. sulcata cognatus* Moskvin et Shimanskaja; (11) *Galeaster dagestanensis* Poslavskaja et Moskvin; (12) *Neoglobator houzeaui* (Cotteau); (13) *N.* aff. *houzeaui* (Cotteau); (14) *N. insolitus* Endelman; (15) *N. panteleevi* Endelman; (16) *N.* sp.; (17) *N. ovalis* Smiser; (18) *N.* cf. *ovalis* Smiser; (19) *Phymosoma* sp.; (20) *Proraster* (*Sanfilippaster*) aff. *geayi* (Cotteau); (21) *Pseudogibbaster tercensis* (Cotteau); (22) *Ps.* cf. *tercensis* (Cotteau); (23) *Studeria crassa* Stoliĉka; (24) *St.* aff. *faberi* (Ravn.); (25) *Trematopygus* aff. *analis* Agassiz.

Crinoids

Kaurtakapy section (Fig. 4)

Usak section (Fig. 8). (1) *Bourgueticrinus danicus* Nielsen; (2) *Bourg.* sp.; (3) *Buchicrinus endelmani* Klikushin; (4) *Buch. paucicirhus* (Nielsen); (5) *Buch. paucicirhus crassus* (Nielsen); (6) *Cainocrinus* sp.; (7) *Denticrinus dentifer* Klikushin; (8) *"Isocrinus" karakalensis* Klikushin; (9) *Isselrcrinus sulzifer* (Eicwald); (10) *Nielsenicrinus fionicus* (Nielsen); (11) *N. varians* Klikushin; (12) *N.* sp.

Bryozoans

Kaurtakapy section (Fig. 4)

Sulukapy section (Fig. 6)

Usak section (Fig. 8). (1) *Coscinopleura angusta minor* Voigt; (2) *Crassicellepora voigti* Berthelsen; (3) *Cystisella fimbriata* (Mokrinskij); (4) *Diplobeisselina nobilis* (Levinsen); (5) *Discovibracella oculata* Voigt; (6) *Dysnoetocella aenigmatica* Voigt; (7) *Florodina atypica* Favorskaja; (8) *Fl. membraniporoides* Voigt; (9) *Frurionella squalida* (Mokrinskij); (10) *Fr.* sp.; (11) *L. faxensis* Levinsen; (12) *"Membranipora" subclavatoides* Voigt; (13) *Micropora*? sp.; (14) *Pavobeisselina oblita* (Kade); (15) *Pliophloea bruennichi* Bertelsen; (16) *P. trabeculifera* Favorskaja; (17) *Psilosecos angustidens* (Levinsen); (18) *Puncturiella* (*Punctiriella*) *cava* Favorskaja; (19) *Puncturiella*? sp.; (20) *Semiescharinella plana* Favorskaja; (21) *Schizmiellopsis anlaltina* (Voigt).

Serpulids

Kaurtakapy section (Fig. 4)

Usak section (Fig. 8). (1) *Crucigera*(?) *insiticia* (Regenhardt); (2) *Neomicrorbis* cf. *sarratus* (Nielsen); (3) *Rotularia* (*Rotularia*) sp.

Corals

Kaurtakapy section (Fig. 4)

Usak section (Fig. 8). (1) *Parisis vertebralis* Hennig; (2) *P. steenstrupi* Nielsen; (3) *Parasmilia* sp.

Selachians (lamnoid sharks)

Usak section (Fig. 8). (1) *Alopias exiqua* Probst; (2) *A.* ex gr. *exiqua* Probst; (3) *A.* ex gr. *latidens* (Leriche); (4) *Echinorynus caspius* Glückman; (5) *Galeocerdo cheganicus* Glückman; (6) *Geleorhinus* ex gr. *latus* Stroms; (7) *Jackelodus trigonalis trigonalis* (Jaeckel); (8) *Lamiostoma menneri* Glückman et Zhelezko; (9) *L.* ex gr. *gracilis* (Le Hon); (10) *Macrorhizodus falcatus* Rog; (11) *M. gigas* Glückman; (13) *Neotidanus* ex gr. *primigensis* Agassiz; (14) *Odontaspis denticulata* Agassiz; (15) *Od. dubia* Agassiz; (16) *Procarchadon sokolovi* (Jackel); (17) *Striatolamia rossica rossica* (Jackel).

Sulukapy section (Fig. 6). (1) *Chararodus* sp.; (2) Jackelotodus spaeri Dartevelle et Cassier; (3) *Odontaspis* sp.; (4) *Striatolamia* aff. *striata* (Winkler).

Microfauna

Kaurtakapy section (Fig. 5)

Sulukapy section (Figs. 6 and 7)

Usak section (Fig. 8)

Smaller foraminifers

Benthic foraminifers (1) *Alabamina wilcoxensis* Toulmin; (2) *Allomorphina halli* Jenning; (3) *Angulogerina wilcoxensis* Cushman et Ponton; (4) *Anomalina danica* (Brotzen); (5) *A. grandis* Vassilenko; (6) *A. granosa* Hantken; (7) *A. fera* Schutzkaja; (8) *A.* ex gr. *welleri* (Plummer); (9) *Bolivina antegressa* Subbotina; (10) *B. missisippiensis* Cushman; (11) *B. simplex* N. Bykova; (12) *Brotzenella acuta* (Plummer); (13) *B. taurica* (Samoilova); (14) *Bulimina granjuscula* Averburg; (15) *Caucasina* schischkinsyae Samoilova; (16) *Cibicides kurganicus* Neckaja; (17) *Cibicidoides clipeatus* Vassilenko; (18) *C. commatus* (Morozova); (19) *C. expertus* (Schutzkaja et Ter-Grigorian); (20) *C. invisus* (Vassilenko); (21) *C. lectus* (Vassilenko); (22) *C. oligocenicus* (Samoilova); (23) *C.* ex gr. *succedens* Brotzen; (24) *C.* ex gr. *ungerianus* (Orbigny); (25) *Clavulina colomi* Hantken; (26) *Coleites reticulosus* Plumer; (27) *Discorbis binkhorsti* (Reuss); (28) *Gavelinella pertusa* (Marsson); (29) *Globorotalites perforatus* Vassilenko; (30) *Gyroidina pontoni* Brotzen; (31) *G. soldanii* (Orbignyi); (32) *Guttulina* (diverse); (33) *Hanzawaia ekblomi* Brotzen; (34) *Heterolepa almaensis* (Samoilova); (35) *H. bratus* (Schutzkaja); (36) *H. hemicompressus* (Morozova); (37) *H.* aff. *pseudoungerianus* (Cushman); (38) *H. sulzensis* (Heer); (39) *Karreria fallax* Rzehak.; (40) *Lenticulina herrmanni* (Andrea); (41) *Melonis dosuralensis* (Chalilov); (42) *Osangularia lens* Brotzen; (43) *Planulina costata* (Hantken);

(44) *Pyramidina crassa* Brotzen; (45) *Pseudogaudryina externa* (Bugrova); (46) *Reussella paleocenica* (Brotzen); (47) *Spiroplectammina azovensis* J. Nikitina; (48) *S. carinata oligocenica* I. Nikitina; (49) *S. spectabilis* Grzybovsky; (50) *S. varianta* Vassilenko; (51) *S. vera* J. Nikitina; (52) *Stensioeina caucasina* Subbotina; (53) *S. whitei* Morozova; (54) *Uvigerina costellata* Morozova; (55) *U. jacksonensis* Cushman; (56) *U. pygmea* Hantken; (57) *Valvulineria ravni* Brotzen.

Planktonic foraminifers (58) larger *Globigerina*; (59) *Acarinina acarinata* Subbotina; (60) *A. bullbrooki* Bolli; (61) *A. djanensis* Schutzkaja; (62) *A. inconstans* Subbotina; (63) *A. marksi* (Martin); (64) *A. rotundimarginata* Subbotina; (65) *A.* ex gr. *rotundimarginata* Subbotina; (66) *A. soldadoensis* (Bronnimann); (67) A. *subsphaerica* Subbotina; (68) *Globigerapsis index* (Finlay); (69) *G. tropicalis* (Blow et Banner); (70) *Globigerina boweri* Bolli; (71) *G. contorta* Schutzkaja; (72) *G. crassa* Schutzkaja; (73) *G. discorda* Kopaevitch; (74) *G. incisa* Hillebrandt; (75) *G. instabilis* Korovina; (76) *G. nana* Chalilov; (77) *G. ratusa* Kopaevitch; (78) *G. subtriloculinoides* Chalilov; (79) *G. triloculinoides* Plummer; (80) *G. turcmenica* Chalilov; (81) *G. varianta* Subbotina; (82) *G. velascoensis* Cushman; (83) *Globoconusa kozlowskii* (Brotzen et Pozaryska); (84) *Globorotalia aequa* Cushman et Renz; (85) *G. aragonensis* Nuttal; (86) *G. caucasica* (Glaessner); (87) *G.* ex gr. *ehrenbergi* (Bolli); (88) *G. marginodentata* Subbotina; (89) *G. subbotinae* Morozova; (90) *Heterochelis gracillima* (Andrea); (91) *Pseudohastigerina* sp.

Larger foraminifers (92) *Daviesina primitiva* Hofker; (93) *D.* ex gr. *archiaci* (Schaub); (94) *Lepidorbitoides mammilata* (Schlumberger); (95) *Nummulites alpina* Douville; (96) *N. anomalis* de la Harpe; (97) *N. atacicus* Leymer; (98) *N. distans* Deshayes; (99) *N. irregularis* Deshayes; (100) *N. murchisoni* Bütneyer; (101) *N. rotularis* Deshayes; (102) *Operculina* ex gr. *parva* Douville et German; (103) *Orbitoides opicula* Schlumberger.

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Evolutionary Relationship of the Eocene Diatom Flora to Some Abiotic Factors

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Abstract—The Eocene evolution of the diatom flora of oceanic and epicontinental sea basins is discussed within the context of some abiotic factors. It was found that from the beginning of the Eocene to the end of the Lutetian, the flora of the Eurosiberian type was widely spread, and the flora of the Atlantic type began to progress from the middle of the early Miocene to reach its prosperity at the end of the Lutetian and the beginning of the Bartonian. The evolution of the flora in basins was controlled by the migration of biogenic elements into the euphotic zone, which in turn, depended upon volcanic activity, transgressions, climate, and lateral and vertical currents. The taxonomic composition of the flora in separate basins is a reflection of a combined effect of global and local factors on evolutionary and migration processes.

Key words: *diatom flora, Eocene, oceanic and epicontinental sea basins, siliceous sediments, volcanism, transgressions.*

Based on the voluminous material obtained in the course of deep sea drilling in ocean floor sediments, a new model of biogenic silica accumulation during the Cenozoic was recently elaborated (Jousé, 1974; Lisitsyn, 1966, 1980; Barron and Baldauf, 1989; Baldauf and Barron, 1990; Baldauf *et al.*, 1992; and others). It was found that diatom algae were the main rock-forming elements.

Prior to these investigations, it was common knowledge, especially among geologists, that low temperature was a principal factor assisting the mass abundance of diatoms. However, current investigations show no direct relationship between the evolution of diatoms and low temperatures. According to Proshkina-Lavrenko (1974), diatoms exist at +1 to +30°C, but the most favorable temperature is +10 to +15°C for cryophilic species and +20 to +28°C for thermophilic species. Among present-day diatoms, rich tropical and subtropical assemblages are recorded in addition to arctic and boreal. During the warm Eocene epoch, the tropical belt of biogenic silica accumulation was formed due to the mass evolution of diatoms. Reserves of Paleogene diatomites, estimated to be 2 billion tons with 0.6 to 9.95×10^6 of unbroken valves per 1 g of diatomite indicate the tremendous productivity of the diatoms which inhabited sea basins in northern Europe (*Kremnistye* ..., 1976).

At present, most researchers have concluded that a sufficient content of nutrients in the euphotic zone is a controlling factor in the diatom vital activity. According to Averina (1968), the mass diatom abundance ($0.1–1.5 \times 10^9$ valves per 1 m^3) in the coastal water of the Atlantic Ocean, near Dakar, occurred at a phosphorus content ranging from 30 to 65 mg/m^3, and a silicon dioxide content from 150 to 200 mg/m^3. Usually, the surface layers of seas and oceans contain no more than 0.5–0.6 mg per 1 litre (Strakhov, 1963).

Silica, an important biogenic element, is involved not only in constructing the shell, but in the metabolism of a diatom cell as well. According to Lisitsyn (1966, 1980), oceans receive the dissolved silicon dioxide from continents as a result of the chemical weathering of rocks, as well as with products of volcanic activity. Whatever their origin, silicic acid and other biogenic elements accumulated in oceans enter the euphotic zone of certain regions due to strong vertical circulation to ensure the vital activity of diversified biota. Lisitsyn believes that the total biogenic siliceous sediments could be formed in the Eocene due to activation of volcanism closely related to geodynamic factors, such as the reorganization of spreading zones, plate motion, and abrupt fluctuations of the sea level against the background of the global transgression.

A harmonious hypothesis has been elaborated for oceans and marginal seas to explain peculiarities in the distribution of respective diatomaceous rocks; however, it is inapplicable to epicontinental sea basins, although there are some works concerning the factors determining the rock-forming role of diatoms.

The present work continues the earlier study (Gleser, 1994) in which the basic stages in the evolution of the Paleogene diatom flora in oceanic and epicontinental sea basins of northern Eurasia were outlined. The objective of this work is to picture the diatom flora evolution in the context of the changing environmental factors throughout the Eocene, one of the most intriguing epochs, as it was the time of the greatest catastro-

phes and changes in the Mesozoic system of the plate arrangement and in the Cenozoic–Recent system of environmental conditions (Lisitsyn, 1980). In addition, it was the time of the widest distribution of diatom algae in inland seas and of their great expansion in oceans, when this group reached the peak of their rock-building role and active evolution during the Paleogene.

The solution of complicated problems on the direct sources of biogenic elements and the impact of local conditions on the action of global factors will help in the interpretation of the reasons for the active diatom life in separate basins during certain intervals of the Eocene.

The diatom flora evolution in inland basins and in oceans during the Eocene was characterized by specific features which resulted from a series of interrelated reasons. Judging from the distribution of diatom algae throughout the Eocene (see figure), it was found that biogenic elements in the World Ocean were derived from different sources.

The mass evolution of diatoms in the West Siberian basin at the end of the Late Cretaceous, throughout the Paleocene (with gaps), in the early Eocene, and in the first half of the middle Eocene could have taken place only as a result of an almost continuous inflow of abundant biogenic elements. In S.B. Shatskii's opinion (1984), they were of volcanic origin. It is conceivable that biogenic elements were transported by currents from the Arctic basin, where they had been brought in by volcanoes situated on the landmass neighboring the East Siberian Sea (Grossgeim, 1975). Currents transported biogenic elements and algae from the West Siberian basin into the Caspian basin and the region west of Mugodzhary through the Turgai strait, where a bit later, in the early and initial middle Eocene, favorable conditions existed for the evolution of the diatom flora almost similar to that of western Siberia.

Lisitsyn *et al.* (1980) believe that accumulation of Eocene biogenic siliceous sediments in the West Siberian basin was confined to the upwelling zones. Judging from the shallow-water flora, the regeneration of biogenic elements occurred in the neritic zone owing to the water circulation induced by solar heating.

In the European paleobasins and on the Canadian coast of the Atlantic Ocean, only short-term outbursts of mass evolution of diatoms are recognized for the Eocene. The peaks in diatom evolution, which were established on the basis of drilling through bottom sediments in the North Sea, coincide with the intensification of volcanic activity (King, 1983).

One of the outbursts of diatom vital activity, which was registered at the very beginning of the Eocene, represents a possible continuation of the active life of the late Paleocene diatoms. Diatoms were probably provided with biogenic elements due to the late Paleocene active volcanism in northwestern Europe (Knox and Harland, 1979). The known diatomite deposit of Denmark (the Moler or Fur Formation) formed at the end of the Paleocene–beginning of the Eocene (Martini, 1977; Perch-Nielsen, 1976). It should be noted that Benda (1972) assigns diatomites to the beginning of the early Eocene, whereas Homann (1991) gives no exact age of flora. More than 130 interlayers of volcanic ash have been recognized in the diatomite sequence. Such a structure of the diatomite sequence is consistent with Murav'ev's idea (1976) that diatomites were formed during periods of less intense explosive volcanism because of high productivity of diatoms in zones of low hydrodynamic activity.

The diatom and silicoflagellate flora of Denmark is similar in composition to that from the middle Volga region near the city of Sengilei (Mt. Granoe Ukho). The similarity of the floras can be explained only by an extensive water exchange. There is good reason to believe that the two floras had a common source of biogenic elements. It is not improbable that at times the flora in the Middle Volga region was also replenished with biogenic elements from the West Siberian basin, insofar as common species of algae were revealed. Distanov (1974, 1976) believes that diatomites in the middle Volga region were formed in river mouths due to the profound chemical weathering of landmasses under the hot humid climate. One of the serious factors contradicting this hypothesis is the exclusively marine character of the algae.

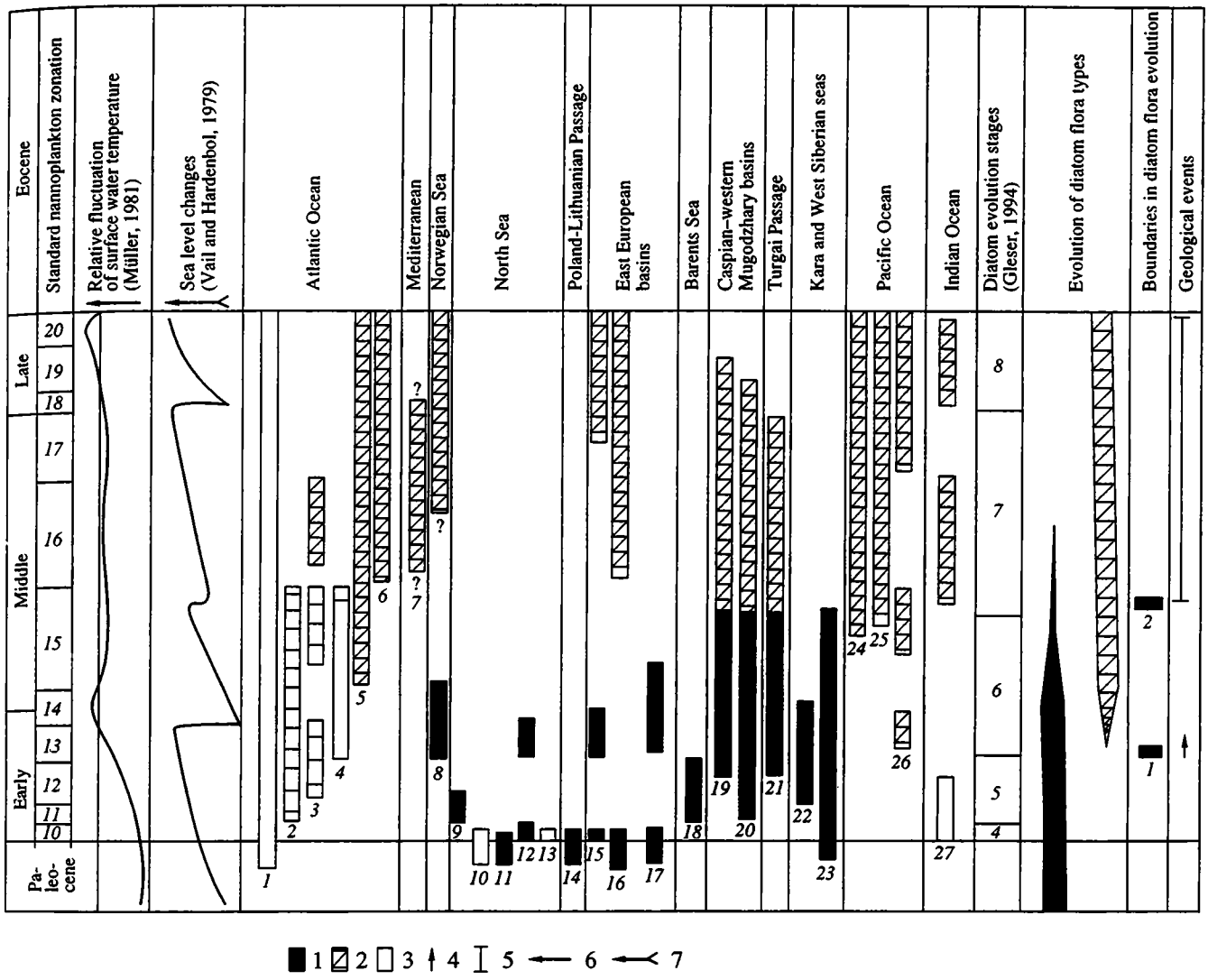
A weaker outburst of diatoms was registered in the Poland–Lithuania strait at the late Paleocene–early Eocene boundary. The diatom flora is of the European type, like that from Denmark and the middle Volga region, as they are very similar in species composition.

The poor preservation of diatom valves excludes definite inferences on the taxonomic composition of the late Paleocene–early Eocene diatom flora revealed on the Canadian coast of the Atlantic Ocean, although Thomas and Gradstein (1981) managed to establish here individual species common to the flora from the North Sea.

Local deposits of small thickness composed of diatom shells are assigned to the very beginning of the Eocene in Belgium (Willems, 1982), northern Germany (Schulz, 1927), the Dnieper–Donets depression, Voronezh anteclise, and Lower Volga region (see figure); later, during a very short time interval, the diatom flora existed in the London basin (Schrubsole and Kitton, 1881). Diatom floras encountered on the East European platform belong to the Eurosiberian type. Other floras are hard to judge, as they either require a revision, or are undefinable because of poor preservation (such as the flora from Belgium).

Starting in the early Eocene, diatom algae inhabit low latitudes of the Atlantic Ocean. In the Gulf of Mexico, the oldest early Eocene diatom flora was found (Fenner, 1984, 1985).

If, prior to this time the flora was of the Eurosiberian type, its appearance in the Gulf of Mexico marked a new epoch of flora of the younger Atlantic type. At first,



Distribution and evolution of diatom floras during the Eocene and changes of some abiotic factors. (*1*) Canadian coast of the Atlantic Ocean (Thomas and Gradstein, 1981); (2) Gulf of Mexico (Fenner, 1984); (*3*) Blake-Bahama (Gombos, 1982; Fenner, 1985; Strel'nikova, 1992); (*4*) Bay of Biscay (data of courtesy of A.P. Padionova); (*5*) Low latitudes of the Atlantic Ocean (Gleser and Jousé, 1974; Fenner, 1984, 1985); (*6*) Falkland Plateau (Fenner, 1984, 1985; Strel'nikova, 1992); (*7*) Mediterranean region, Israel (Ehrlich and Meshkovitz, 1982); (*8*) Norwegian Sea (Schrader and Fenner, 1976; Bukry, 1976; *Istoriya* ..., 1979); (*9*) London basin (Schrubsole and Kitton, 1881); (*10*) North Sea (King, 1983); (*11*) Denmark (Benda, 1972; Homann, 1991); (*12*) Northern Germany (Schulz, 1927; Benda, 1965); (*13*) Belgium (Willems, 1982); (*14*) Kaliningrad region (Strel'nikova *et al.*, 1978); (*15*) Dnieper–Donets depression (Gleser *et al.*, 1965; Gleser and Sheshukova-Poretskaya, 1968, 1969; Ol'shtynskaya, 1977, 1978; unpublished materials of the author); (*16*) Voronezh anteclise (Palatnaya, 1976; Ol'shtynskaya, 1976; Strel'nikova, 1992; unpublished materials of the author); (*17*) Lower and Middle Volga region (Gleser, 1966, 1993; Gleser *et al.*, 1987); (*18*) Barents Sea near Franz Jozef Land (Grunow, 1984); (*19*) Caspian basin (unpublished materials of the author); (*20*) Region west of Mugodzhary (Gleser, 1967; Segedin, 1972; unpublished materials of the author); (*21*) Turgai depression (Gleser, 1966, 1967, 1970); (*22*) Kara Sea (Gleser and Stepanova, 1993); (*23*) West Siberian lowland (Strel'nikova, 1960; Krotov and Shibkova, 1961; Rubina and Droznes, 1961; Paramonova, 1964; Gleser, 1966, 1970, 1978, 1986; unpublished materials of the author); (*24*) Northwestern Pacific Sheshukova-Poretskaya, 1967; Oreshkina, 1982; Gleser *et al.*, 1986; Pushkar', 1987); (*25*) Pacific Ocean, low latitudes (Fenner, 1980, 1985; Barron *et al.*, 1984); (*26*) Southern Pacific (Hajós, 1976; Strel'nikova, 1992); (*27*) Indian Ocean (Fenner and Mikkelsen, 1990; Fourtanier, 1991).

(1) Flora of the Eurosiberian type; (2) flora of the Atlantic type; (3) flora of the undetermined type; (4) conjunction of the West Siberian sea with Norwegian Sea through the Arctic basin; (5) extensive water exchange between the Atlantic Ocean with epicontinental sea basins of northern Eurasia and Norwegian Sea; (6) cooling events; (7) transgressions.

each of these floras evolved in an original way. The oceanic flora of the middle part of the early Eocene is rich in species of the genus *Craspedodiscus*, whereas in epicontinental sea basins, the species of the genus *Coscinodiscus* prevail. The number of species of this eurythermal genus gradually grew from 11 to 27%, whereas the number of thermophilic species of the family Biddulphiaceae decreased from 59 to 28%. Such a change in the taxonomic composition proceeded concurrently with a gradual decrease in the surface water temperature during the early Eocene (Müller, 1981).

From the latest Paleocene to the end of the middle Eocene, the diatom flora existed in different regions of the Indian Ocean. No changes in its taxonomic composition were found here, like in other paleobasins at the Paleocene–Eocene boundary (Fourtanier, 1991).

This can be reasonably explained by the absence of abrupt fluctuations of the sea level (Vail and Hardenbol, 1979) and temperature at the boundary between the two epochs. There are individual common species with the flora of the Eurosiberian type inherited mainly from the late Paleocene flora; the flora of the Indian Ocean evolved in an original way.

The terminal early and initial middle Eocene intervals show fundamental changes in the evolution of the diatom flora and environmental conditions.

A new short-term outburst of diatoms occurs in the north of the German basin; diatoms seem to start inhabiting the Norwegian Sea. After a break, they appear again in the Ul'yanovsk bay of the middle Volga region. Their distribution areas widen in the West Siberian basin, Caspian region, Turgai strait, and in low latitudes of the Atlantic Ocean. Diatoms also appear in the North Pacific (see figure).

Across the early–middle Eocene boundary, the prevalence of species of the genus *Coscinodiscus* gave way to the domination of species from the genus *Pyxilia*, which were better adapted to more dynamic hydrologic conditions. A gradual widening of distribution areas of the diatom flora against the background of the growing transgression caused an exchange of individual species between the Eurasian and Atlantic floras. In the West Siberian sea, Turgai strait, and on the shelf west of Mugodzhary, the taxonomic composition of diatom floras gradually changed; the enrichment of diatomites in terrigenous material and glauconite, as well as replacement of them by clays, were equally gradual. In the Norwegian Sea and northern German basin, the algal flora changed drastically: nannoplankton replaced diatoms and silicoflagellates, and the carbonate sedimentation (NP 12 Zone) abruptly gave way to siliceous sedimentation: Unter-Eozän 4 in northern Germany (Martini, 1981), Unit 2 in the Norwegian Sea, Site 343 (Müller, 1976).

Based on the comparative analysis of diatom and silicoflagellate floras from the Norwegian Sea and those from the Atlantic Ocean and northern Eurasia, we can infer that the Norwegian Sea became connected with the Arctic basin in the early–middle Eocene boundary epoch.

Diatom and silicoflagellate assemblages of the Norwegian Sea, which were distinguished in carbonate-free sediments at the level higher than the lower–middle Eocene boundary (Site 343), are certainly of the Eurosiberian type. The presence of *Pyxilla oligocaenica* Jousé, the species unknown in the Atlantic floras, is very characteristic here. In the systematic taxonomic and species abundance, the silicoflagellate assemblage is identical to that from the upper part of the *Dictyocha secta* Zone, which is traceable from the Red Sea to the Caspian region and has not yet been found in the Atlantic (Martini and Müller, 1976). It should be noted that the lower part of Unit 2 (9.5 m), occurring between the early–middle Eocene boundary and NP 12 Zone, has not been studied.

Still more significant variations in the taxonomic composition and distribution areas of diatom flora took place parallel to considerable environmental changes across the Lutetian–Bartonian boundary.

According to available data (Barron and Baldauf, 1989; Baldauf and Barron, 1990), the peak of the diatom algae outburst and remarkably high rates of biogenic sedimentation in the middle Eocene resulted from a mass inflow of biogenic elements into the euphotic zone. The polar cooling caused the growth of the temperature difference between the surface and deep water, the appearance of thermocline, and a higher latitudianl thermal gradient that resulted in the vertical circulation of deep water in certain ocean regions. In the second half of the middle Eocene, the rock-building role of diatoms was manifested in the upwelling zones, the tropical belt of the biogenic silica accumulation was formed, and diatomaceous sediments were deposited in coastal regions of California, Kamchatka, Peru, and New Zealand (Baldauf and Barron, 1990).

The more or less constant warming, mass inflow of nutrients, and high sea level created the most favorable conditions for the evolutionary outburst in the high-rank taxa of the class Centriphyceae, as well as for a fast and wide distribution of numerous young species.

The diatom flora of the Eurosiberian type, which dominated from the end of the Late Cretaceous, was replaced by the Atlantic-type flora.

Most of species typical of this flora (*Pyxidicula moelleri* Grun., *Stephanopyxis marginata* Grun., *S. megapora* Grun., *Pseudopodosira pileiformis* Jousé, *P. bella* Possn. et Gles., *Hyalodiscus radiatus* Grun., *Anuloplicata ornata* (Grun.) Gles., *Coscinodiscus decrescens* Jousé *C. payeri* Grun., *C. polyactis* A. Cl., *Pseudotriceratium chenevieri* (Meist) Gles., *Hemiaulus elegans* var. *intermedia* Grun., *H. polymorphus* Grun. var. *frigida* Grun., *Gomotnecium odontella* Ehr., *Trinacria excavata* Heib., *T. regina* Heib., *Pyxilla gracilis* Temp. et Forti, and *P. oligocaenica* Jousé) gradually gave way to species from ocean basins; few of them migrated into oceans. Representatives of the

Atlantic-type flora inhabiting northern sea basins are mainly species of a wide geographic distribution. They comprise *Coscinodiscus denarius* A.S., *C. decrescens* Grev., *Cosmiodiscus senarius* (A.S.) Gles. et Olscht., *Craspedodiscus oblongus* (Grev.) Hanna, *Cristodiscus succinctus* (Sheshuk. et Gles.) Gles. et Olscht., *Brightwellia coronata* Ralfs, *Pseudotriceratium coscinoides* Gr. et St. Gles., *Pronia barbadensis* Grev., *Hemiaulus polycystinorum* Ehr., *Monobrachia simplex* Schrader, *Lisitzinia inconspiqua* (Grev.) Gles., var. *inconspiqua* et var. *trilobata* (Fenner) Gles., *Sheshukovia condecora* (Brightw) Gles., *Pyxilla johnsoniana* (Grev.), and others. Species of this group cannot be considered as indicating the climatic warming in the middle Eocene. The appearance of some of them seems to be related to the sea level fluctuations. The first appearance of the oceanic species *Craspedodiscus oblongus* (Grev.) Hanna among the diatom flora of the Caspian region coincides with the transgression maximum at the end of the early Eocene, then it disappears probably because of the regression at the beginning of the middle Eocene. During the sea level rise at the end of the Lutetian–Bartonian, this species becomes a constant component of the flora; horizons with this species in the West Siberian sea correspond to the end of the Lutetian. Its appearance levels in different oceanic regions (Fenner, 1985; McLean and Barron, 1988; Strel'nikova, 1992) are also controlled by the depth of basins and are not synchronous.

Oceanic, highly specialized planktonic species of the genera *Asterolampra* and *Rhylandsia* of the family Asterolampraceae, as well as *Azpeitia* of the family Azpeitiaceae, which are characteristic of the Atlantic-type flora, hardly penetrated into epicontinental basins of northern Eurasia, like the group of tropical species *Craspedodiscus undulatus* Gombos, *Hemiaulus alatus* Grev., and *H. gondolaeformis* Fenner. The latter are associated with *Lisitzinia kanayae* (Fenner) Gles., the species typical of the low and middle latitudes.

According to Fenner (1985), the climatic differentiation in the second half of the middle Eocene resulted in the formation of two major paleobiogeographic provinces: the area of low latitudes and the area of high latitudes of the southern hemisphere.

Expanding in oceans, the distribution areas of the flora decreased in northern Eurasia. At the end of the Bartonian, the connection of the West Siberian basin with the Arctic region ceased due to uplifts in the north (Shatskii, 1984). Even earlier, probably at the end of the Lutetian, the cessation of inflow of biogenic elements into the West Siberian sea drastically affected diatoms and silicoflagellates. However, diatom algae still existed in the Turgai strait, Caspian basin and the region west of Mugodzhary, and migrated into basins on the territory of the Voronezh syneclise and possibly the Dnieper–Donets depression. It seems that in spite of the ceased inflow of biogenic elements from the north, the diatom vital activity was supplied with nutrients which were transported through the expanding (due to trangression) Poland–Lithuanian and southern straits of the Tethys. Elements of the Atlantic type had a significant role in the flora of these basins.

Most likely, when the diatom flora penetrated from the Atlantic Ocean into the Norwegian Sea, an intense exchange of water, species, and biogenic elements between these two basins was set in.

The transition from the Bartonian to late Eocene flora was very gradual. Neither in oceans (Fenner, 1985), nor in epicontinental basins of northern Eurasia have the pronounced distinctions in the taxonomic composition of the diatom flora across the middle–late Eocene boundary been found. In oceans, the onset of a new evolutionary stage of the diatom flora corresponds to the middle of the late Eocene. In the Dnieper–Donets basin, where the late Eocene flora is represented more completely, the second stage of its evolution has not been recognized. It should be pointed out that in the late Eocene, the species of different young genera and families of the class Pennatophyceae become more actively involved in the formation of benthic associations. This class is characterized by a more diverse taxonomic composition in New Zealand (Oamaru) (Schrader, 1969), whereas in the Dnieper–Donets basin it is much poorer. It may be suggested that the enhancement of the taxonomic diversity of benthic associations due to the intensified evolution of the class Permatophyceae could be related to the expansion of the littoral zone because of the sea level fall at the beginning of the late Eocene.

Thus, the current attempt to elaborate a general model for the diatom flora evolution in oceans and epicontinental sea basins made it possible to contemplate relation between separate moments of the flora evolution and abiotic environmental transformations throughout the Eocene. Two types of the flora have been recognized: the older flora, assigned to the Eurosiberian type, and the younger flora, conventionally attributed to the Atlantic type.

From the early Eocene to the middle level of its middle subseries, the diatom flora of the Eurosiberian type existed in epicontinental sea basins, and gradually gave way to younger flora of the Atlantic type.

A significant stage in the evolution of the diatom flora of the Eurosiberian type occurred at the end of the early Eocene. The changing character and ranges of the flora reflect the changes in the hydrodynamic and hydrochemical regime of epicontinental sea basins in northern Eurasia. These transformations seem to be caused by the emerging wide connection between the West Siberian basin and the Norwegian Sea through the Arctic basin.

The second important stage of the flora evolution in oceanic and epicontinental basins corresponded to the end of the Lutetian–beginning of the Bartonian. The most favorable conditions for evolution of the Atlantic type flora appeared in oceanic upwelling zones formed due to polar cooling, which changed the circulation of

deep and surface water. At the same time, the ceased inflow of biogenic elements in the West Siberian basin, and further in the Turgai strait, the Caspian basin and the region west of Mugodzhary was a catastrophic event for the flora of the Eurosiberian type. An intensive water exchange between oceans as well as between the Atlantic ocean and the Norwegian and epicontinental seas of northern Eurasia, which was initiated by the transgression in the middle and earliest late Eocene time, resulted in the extensive migration of the eurybiontic elements of Atlantic-type flora.

The regression of the World Ocean, which started in the second half of the late Eocene, as well as the enhanced climatic zonation, gave rise to specific features of floras in different oceanic regions, and triggered the development of benthonic assemblages. Local conditions brought their own corrections into the effect of global factors; they controlled the character of flora and its distribution in separate paleobasins.

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Pliocene–Quaternary Biostratigraphy and Planktonic Foraminifera Evolution in the North Pacific

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Abstract—In analysis of the stratigraphic distribution of planktonic foraminifers in the Pliocene–Lower Quaternary sediments of the North Pacific, and research of the space-time distribution dynamics of 60 species at four levels corresponding to the foraminiferal zones N19, N20, N21, and N22 revealed, significant evolutionary changes at that time were revealed. Twenty-five species became extinct and ten species appeared. Since the early Pliocene, the foraminifer diversity decreased from 53 to 34 species. Three groups of species were revealed: (1) short-lived species; (2) species who reduced and extended distribution areas and abundances, and (3) species whose distribution patterns remained relatively stable. It was shown that the reduction of distribution areas proceeded from the periphery toward the center (i.e., toward the areas with the highest abundances) and, in contrast, distribution areas of the newly appearing and evolving species expanded from their center toward the periphery.

Key words: *planktonic foraminifers, Pliocene–Quaternary sediments, biostratigraphic distribution, datum plane, distribution areas, diversity, biogeographic provinces, extinction, evolutionary origin.*

INTRODUCTION

The study of distribution patterns of planktonic foraminifer species in oceanic sediments is the first important step in elaborating various biostratigraphic scales and defining paleoceanographic changes. Planktonic foraminifers provide a unique opportunity to carry out these investigations because of the fast evolutionary changes in this group of microorganisms during relatively short time intervals, dependence of their distribution on climatic zonality in the oceans, and sensitivity to the environments in surficial waters (Barash, 1988).

Recently, the stratigraphic subdivision of the Pliocene–Quaternary sediments of the World Ocean based on planktonic foraminifers was well substantiated: many levels of species extinction and appearance (datum planes) and other events in microplankton evolution were established. Based on these events, the criteria for defining boundaries of biozones and other stratigraphic units were elaborated. About 40 datum planes in the Pliocene–Quaternary sediments of the World Ocean are noted in many publications (Saito, 1977; Thompson and Sciarillo, 1978; Berggren, 1984). Still, the position of these datum planes in the particular sections can depend on local climatic and facies peculiarities, while not necessary coinciding with the level of the real evolutionary appearance or extinction of species (Dowsett, 1988; and others). This renders detailed biostratigraphic studies on the evolution of planktonic foraminifers in particular oceanic areas and regional peculiarities in the stratigraphic distribution of various species necessary. Several summarizing works were dedicated to the stratigraphic aspects of foraminifer distribution in the Neogene–Quaternary sediments of the Pacific (Parker, 1967; Kennett and Srinivasan, 1984; Bolli and Saunders, 1985). In many publications, special attention was paid to the problems of evolutionary changes in this microplankton group in the Southern Hemisphere (Kennett and Srinivasan, 1983; Krasheninnikov and Basov, 1986; Jenkins, 1992a; Jenkins, 1992b; and others). Similar data on planktonic foraminifers in the Northern Hemisphere are less abundant (Keller, 1978; Keller, 1979; and others).

The analysis of paleogeographic regularities in a planktonic foraminifer distribution allows their assemblages to be used as indicators of paleoclimatic and paleoceanographic environments. This is evident from the confinement of foraminiferal assemblages, which are characterized by a certain number of species and their ratio to particular water masses and climatic zones. When reconstructing paleoceanographic changes based on biogeographic data, information on the confinement of fossil species to particular bioprovinces is used. The distribution characteristics of recent foraminiferal species with well-known ecological features can be ascribed to their fossil counterparts (Barash, 1988). Thus, the concept on the climatic preference of fossil species and the position of foraminifer assemblages in past bioprovinces is formed. The recognition of provinces whose climatic characteristics are predetermined by their geographic setting allows us to elaborate various paleoceanographic reconstructions (Blyum *et al.*, 1987). In the Pacific, similar paleobiogeographic studies were performed by many research-

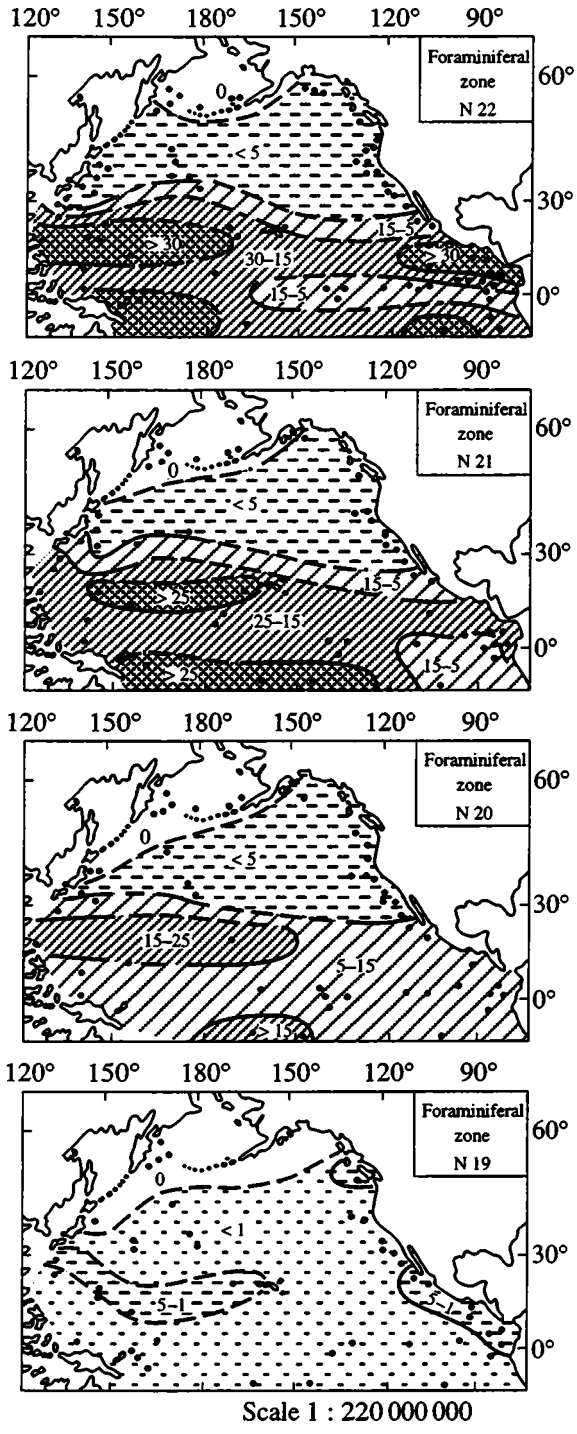


Fig. 1. Maps of distribution areas and percentage of species *Globigerinoides ruber* for time intervals of N19, N20, N21, and N22 foraminiferal zones in Pliocene–Lower Quaternary sediments of the North Pacific.

ers (Blyum and Sokolova, 1989; Blyum *et al.*, 1989; Jenkins, 1992a). The distinguishing feature of the present study is that we used, when defining the peculiarities of the biogeographic distribution of planktonic foraminifers, the quantative characteristics of their distribution areas, and this approach provides more reliable and detailed reconstructions.

MATERIALS AND METHODS

Five hundred and seventy samples from the Pliocene–Lower Quaternary sediments drilled in the North Pacific at the latitudes between 15°S and 60°N were available for the analysis. Biogeographic studies were performed for four time intervals, which, according to the newly adopted Pliocene–Quaternary stratigraphic zonation, correspond to foraminiferal zones N19 (100 samples), N20 (122 samples), N21 (155 samples), and N22 (193 samples). Data on zone volumes and age estimates of their boundaries are taken from combined zonal schemes (Krasheninnikov, 1978; Krasheninnikov and Basov, 1986). When specifying the sample ages, we used data on the stratigraphic distribution of planktonic foraminifer species (Kennett and Srinivasan, 1983; Bolli and Saunders, 1985) and datum planes (Berggren *et al.*, 1985; and others).

Approximately 100 foraminiferal species have been defined in the Pliocene–Quaternary sediments throughout the global ocean by various researchers. In the studied samples, which contained foraminifer shells of variable degrees of preservation, we found and identified (based on publications (Bolli and Saunders, 1985; Stainforth *et al.*, 1975; and others)) 65 species and morphological varieties. In doing so, we adopted the taxonomic classification of Kennett and Srinivasan (1983). Some relative species and subspecies were united with the type forms: *Globigerinoides triloba* and *G. immaturus* with *G. quadrilobatus; Globigerina* aff. *woodi* and *G. apertura* with *G. decoraperta*; and *Pulleniatina precursor* with *P. primalis*. When interpreting paleobiogeographic characteristics, we considered some genetically related species together: *Globorotalia conomiozea* and *G. conoidea; Orbulina universa, O. suturalis*, and *O. bilobata*; *Neogloboquadrina dutertrei* and *N. dutertrei blowi*; and *Globigerinella calida* and *G. precalida*.

The quantative analysis of species in fraction >0.1 mm was performed and the percentage of each of the species was determined in all the samples studied. In doing so, we excluded, in accordance with the method by Ivanova (1988), *Globiberinita glutinata* from the calculations. Using the obtained results, we compiled distribution maps of 59 species and varieties of planktonic foraminifers for the considered time intervals. Figure 1 shows, as an example, the schematic distribution map of *Globigerinoides ruber*. In addition to our own materials, we also used data on planktonic foraminifer distribution in the Pliocene–Lower Quaternary sediments cited in the Deep Sea Drilling Project and Ocean Drilling Program reports, among other works (Keller, 1978, 1979; and others). Using the adopted method, we recalculated quantative data from these publications and correlated them with our own results. Semiquantative and qualitative data were used

for specifying the boundaries of species distribution areas.

Further, based on these maps, an analysis of the latitudinal distribution of different species was performed, and schemes showing the distribution area changes with time were compiled. The average percentage was calculated for widely distributed species in every latitudinal belt. Note that, in contrast to this approach, the highest species contents were considered in similar studies on recent planktonic foraminifers (Barash, 1974). The use of average contents in analyzing fossil foraminifers is predetermined by the duration of the studied time intervals (about 1 m.y.), during which significant variations in species contents occur because of the multiple changes in environmental parameters, evolutionary peculiarities of different species, and influence of selective dissolution. The highest contents were used only for compiling the distribution areas of rare species.

The Pliocene–Early Quaternary planktonic foraminifer assemblages in the North Pacific are characterized by their varying degrees of preservation. The influence of dissolution generally increases toward the higher latitudes and in the west–east direction. This corresponds to the distribution patterns of planktonic foraminifers in recent sediments of the North Pacific, which are controlled by the position of foraminiferal lysocline (Belyaeva, 1980; and others). Taking these peculiarities into consideration, we used, when compiling schemes of distribution areas, mainly quantative data obtained along the meridional profiles throughout the central and northwestern parts of the ocean. For high-latitude species, data from the northeastern Pacific were also considered.

RESULTS AND DISCUSSION

The studies performed allowed us to define the regularities in the biogeographic distribution of foraminiferal species in the Pliocene–Lower Quaternary sediments of the North Pacific and reveal some of the peculiarities in their evolution. Based on the analysis of the stratigraphic ranges of the observed species, we compiled a scheme of their stratigraphic distribution (Fig. 2). One of the earliest stratigraphic schemes compiled for the equatorial-tropical realm of the Indian and Pacific oceans (Parker, 1967) shows stratigraphic ranges of 34 Pliocene–Quaternary species. Most of them do not coincide with those obtained from our materials; sometimes, the discrepancy between data is equal to the volume of the foraminiferal zone. According to our data, species *Globigerinoides ruber, G. obliquus, Neogloboquadrina dutertrei, Globigerina nepenthes*, and others have a wider stratigraphic range, whereas *Globigerinoides fistulosus, Pulleniatina spectabilis, Globoquadrina venezuelana* and others are characterized by a narrower range. The later stratigraphic schemes show no essential discrepancies with ours, although some differences can be noted. For instance, our scheme includes species *Globigerina incisa, Globoquadrina pseudofoliata, Neogloboquadrina pseudoopima*, and *Globorotalia pumilio*, which are absent from the scheme by Kennett and Srinivasan (1983) compiled for Neogene deposits of the entire Pacific, but based mainly on studies in the southwestern Pacific. Some species, according to our data, have a wider stratigraphic range: *Globigerina decoraperta, G. rubescens, G. umbilicata, Neogloboquadrina dutertrei, N. acostaensis, Globorotalia ungulata, G. limbata*, and *G. pertenuis*. In the scheme by Bolli and Saunders (1985) proposed for lower latitudes and based mostly on foraminiferal data from the Atlantic Ocean, *Globoquadrina pseudofoliata* is absent, according to our data, however, this species should be considered as the equatorial-tropical form. In our scheme, a wider stratigraphic range is characteristic of the species *Neogloboquadrina dutertrei, N. acostaensis, Pulleniatina obliquiloculata, P. primalis, Globorotalia inflata*, and *G. crassaformis*, while a narrower range is typical of the species *Globigerina nepenthes, Globigerinoides obliquus, G. extremus, Neoqloboquadrina humerosa, Globorotalia ungulata, G. exilis, G. pertenuis*, and *Globigerinella aequilateralis*. These differences can probably be explained by peculiarities in the evolution and ecology of these species, which result in their different stratigraphic ranges in various regions of the global ocean.

The planktonic foraminifer diversity varied during the considered time interval, from 52–53 species in the early Pliocene to 34 species in the Eopleistocene (Fig. 2).

According to Wei and Kennett (1986), who analyzed the evolution of planktonic foraminifers through the Neogene using the data on phylogenetic lineages (Kennett and Srinivasan, 1983), the taxonomic diversity decreased during last 5 m.y. from 59 late Miocene species to 44 late Quaternary species which could have resulted from the acceleration of paleoceanographic oscillations. Dating of evolutionary changes in some phylogenetic lineages of planktonic foraminifers (for instance, in the *Globorotalia inflata, G. tumida, Sphaeroidinella dehiscens*, and Globorotalia truncatulinoides lineages) during the late Neogene also evidence that they occurred almost synchronously with major paleoceanographic events (Malmgren and Berggren, 1987).

Wei and Kennett (1986) showed that the Neogene evolutionary rates of planktonic foraminifers (speciation, extinction, etc.) in three climatic zones (tropical, temperate, and transitional) are similar, although in their opinion, the number of evolutionary events in the tropical zone is always higher because of the higher diversity there. According to our data, during the Pliocene–early Quaternary (N19 to N22 foraminiferal zones), 25 species (13 equatorial-tropical + five tropical + three subtropical) became extinct in the North Pacific and ten species (three + two + five, respectively) appeared. Thus, the taxonomic diversity decreased

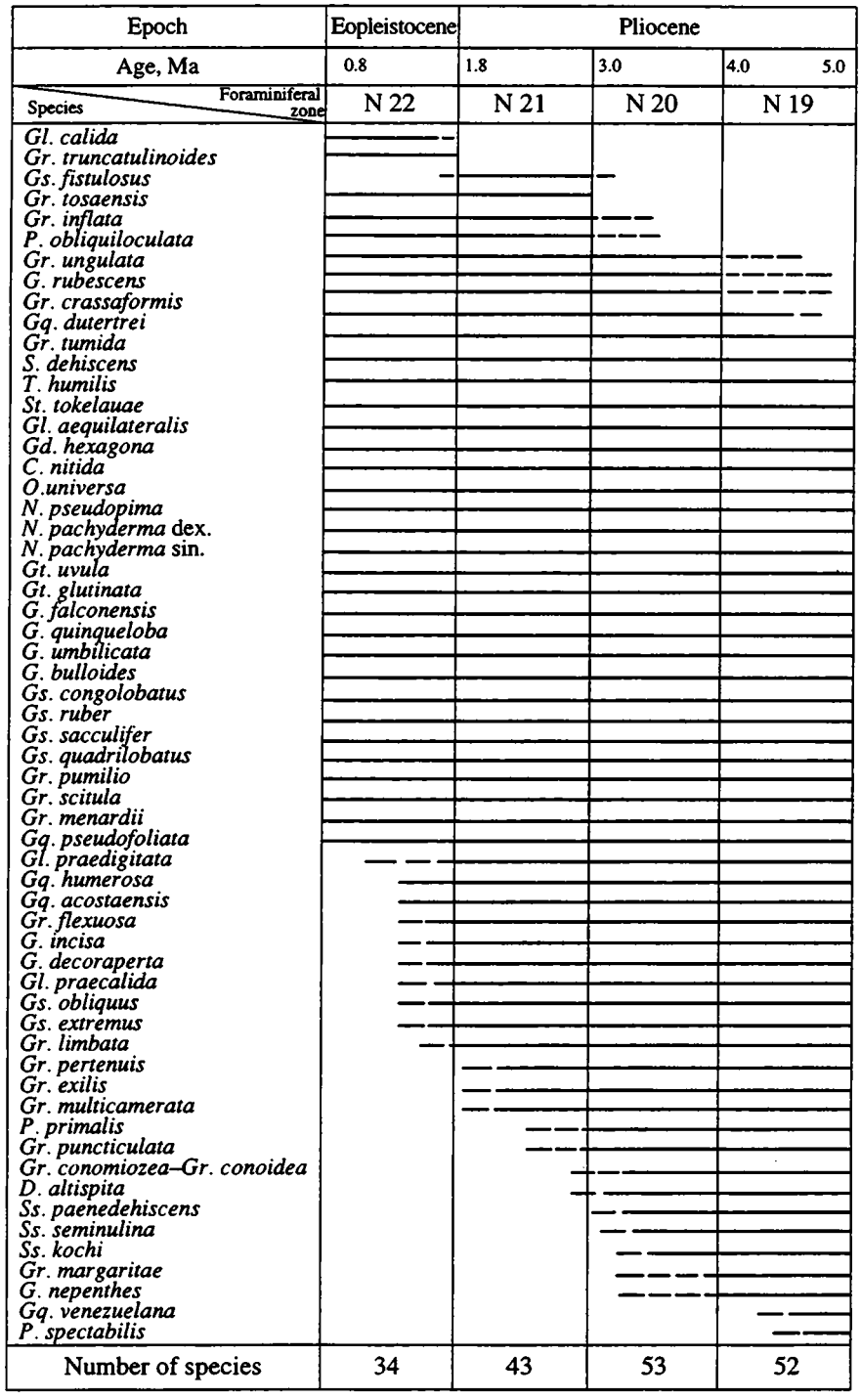


Fig. 2. Stratigraphic distribution of planktonic foraminifer species in Pliocene–Lower Quaternary sediments of the North Pacific.

mainly at the expense of the extinction of equatorial-tropical and tropical species, whereas the most subtropical species continued to evolve. No single species appeared during the considered interval in the temperate belt.

Other peculiarities in planktonic foraminifer evolution were also defined from paleobiogeographic data in the intervals studied. Results of these studies are represented in a series of schemes, which reflect the space–time changes in distribution areas of the species (Fig. 3).

Having a large set of new quantative data, we considered for every species not only the distribution area limits, but also the changes in species concentrations within these areas. Taking into consideration the con-

tents of species in areas of their highest concentrations (see Fig. 3), we defined four groups of Pliocene–Lower Quaternary species differing in abundance: (1) abundant species with concentrations of more than 40% (*Neogloboquadrina pachyderma* sinistral, *Globigerinoides ruber*, and others), (2) common species (40–15%; *Globigerinoides obliquus, G. quadrilobatus*, and others), (3) scanty species (15–5%; *Sphaeroidinellopsis seminulina, Globigerina quinqueloba, Globorotalia foliata*, and others), and (4) rare species (<5%; *Candeina nitida, Globorotalia multicamerata, Globigerinita uvula*, and others).

Species *Globigerinita glutinata*, by our data, is widespread throughout all considered latitudes, and its highest concentrations make up 50% of the entire foraminiferal assemblage of the equatorial and temperate latitudes (Fig. 3, *59*). Many researchers consider this species to be of the recent cosmopolitan form, which shows no distinct relation to particular latitudes and inhabits waters with a wide temperature range. Therefore, we excluded *G. glutinata* from our calculations.

According to spatial distribution, the individual species can be united into three groups, which show the highest concentrations respectively in the lower (0°–25°N), middle (25°–40°N), and higher (40°–60°N) latitudes. According to preliminary analysis, these groups correspond to three major biogeographic provinces of planktonic foraminifers in the World Ocean: equatorial-tropical, subtropical, and temperate-subpolar ones. More detailed paleobiogeographic zonation will be possible after the compilation of maps of the types of Pliocene–Lower Quaternary foraminiferal thanathocoenoses, which is planned to be done in the future. In the southwestern Pacific, Jenkins (1992a) defined five Pliocene provinces, which correspond to the recent ones outlined by Be (1977). As these scientists considered only the qualitative distribution of planktonic foraminifers, the province boundaries were drawn according to the latitudinal limits of species distribution areas which gives no indication of the entire assemblage composition and quantative proportions of species in the particular province. We have such data, and in further studies, they will form the basis for biogeographic zonation.

The analysis of distribution area dynamics (Fig. 3) allows some features of planktonic foraminifer evolution in the North Pacific during the Pliocene–Early Quaternary to be outlined. Using this information, we identified several types of distribution areas differing in dynamics:

(1) Distribution areas of short-lived species, which appeared and became extinct during the considered time interval (Fig. 4, *I*). These are *Globorotalia margaritae, Globigerinoides fustulosus*, and *Pulleniatina spectabilis* (Fig. 3, *50, 11*, and *24*, respectively). These species, of low or very low abundance, were confined to the narrow latitudinal belt, and characterized by a short stratigraphic range within the limits of one or two foraminiferal zones.

(2) Distribution areas of species, which reduced their distribution during the Pliocene–Early Quaternary (Fig. 4, *II A* and *B*). There are two groups of species, which passed from the older pre-Pliocene strata and showed either common, or low abundances in the early Pliocene. The first group (A) includes species, which became gradually less abundant during the Pliocene-Early Quaternary and evolved, reducing their distribution areas from the periphery toward the areas of highest concentrations. These are *Globigerina incisa, Globigerinoides extremus, Globorotalia limbata, G. tosaensis, G. pumilio*, and *Globoquadrina pseudofoliata* (Fig. 3, *8, 14, 34, 43, 48*, and *51*, respectively). The other group (B) consists of species displaying similar evolutionary trends, however, they became extinct during the considered time interval: *Globigerina nepenthes, G. altispira, Globigerinoides obliquus, Sphaerodinelopsis seminulina, S. kochi*, and *Globorotalia conomiozea–G. conoidea* group (Fig. 3, *5, 15, 19, 20, 47*, and *53*, respectively). Rare species *Globorotalia multicamerata, G. exilis, G. pertenuis*, and *Globoquadrina venezuelana* (Fig. 3, *35* to *37* and *52*, respectively) became extinct without changes in their distribution areas. *Globoquadrina pseudofoliata* and *Globorotalia tosaensis*, whose distribution areas were reduced during the Pliocene–Early Quaternary, became extinct somewhat later, in the N23 foraminiferal zone (Thompson and Sciarillo, 1978).

(3) Distribution areas of species, whose distribution widened during the Pliocene–Early Quaternary (Fig. 4, *III A* and *B*). Two types of extending distribution areas can be distinguished. The first one (A) is represented by older species, which experienced maximum flourishing in the Pliocene. *Globigerinoides ruber*, known since the middle Miocene, became common in the Pliocene and abundant in the Eopleistocene (Fig. 1). The pre-Pliocene species *Globigerinella aequilateralis, Globorotalia menardii*, and *G. scitula* also somewhat extended their distribution areas and became more abundant (Fig. 3, *31, 33*, and *49*, respectively).

The second type (*B*) is characteristic of species, which appeared in the Pliocene and evolved through the Pliocene–Quaternary. In the Pliocene–early Quaternary, these species became either abundant (*Globigerina rubescens*), or common (*Globorotalia crassaformis*, and *Neogloboquadrina dutertrei*); others remained scanty (*Sphaeroidinella dehiscens* and *Globorotalia truncatulinoides*) (Fig. 3, *7, 46, 28, 17, 41*, and *42*, respectively).

(4) Distribution areas of genetically related species, which inherited distribution patterns from older species (Fig. 4, *IV*). Species of the phyletic lineages *Pulleniatina primalis–P. obliquiloculata* (Fig. 3, *22* and *23*, respectively), *Globorotalia puncticulata–G. inflata* (Fig. 3, *45* and *44*, respectively), *Globigerina decoraperta–G. rebescens* (Fig. 3, *6* and *7*, respectively), and

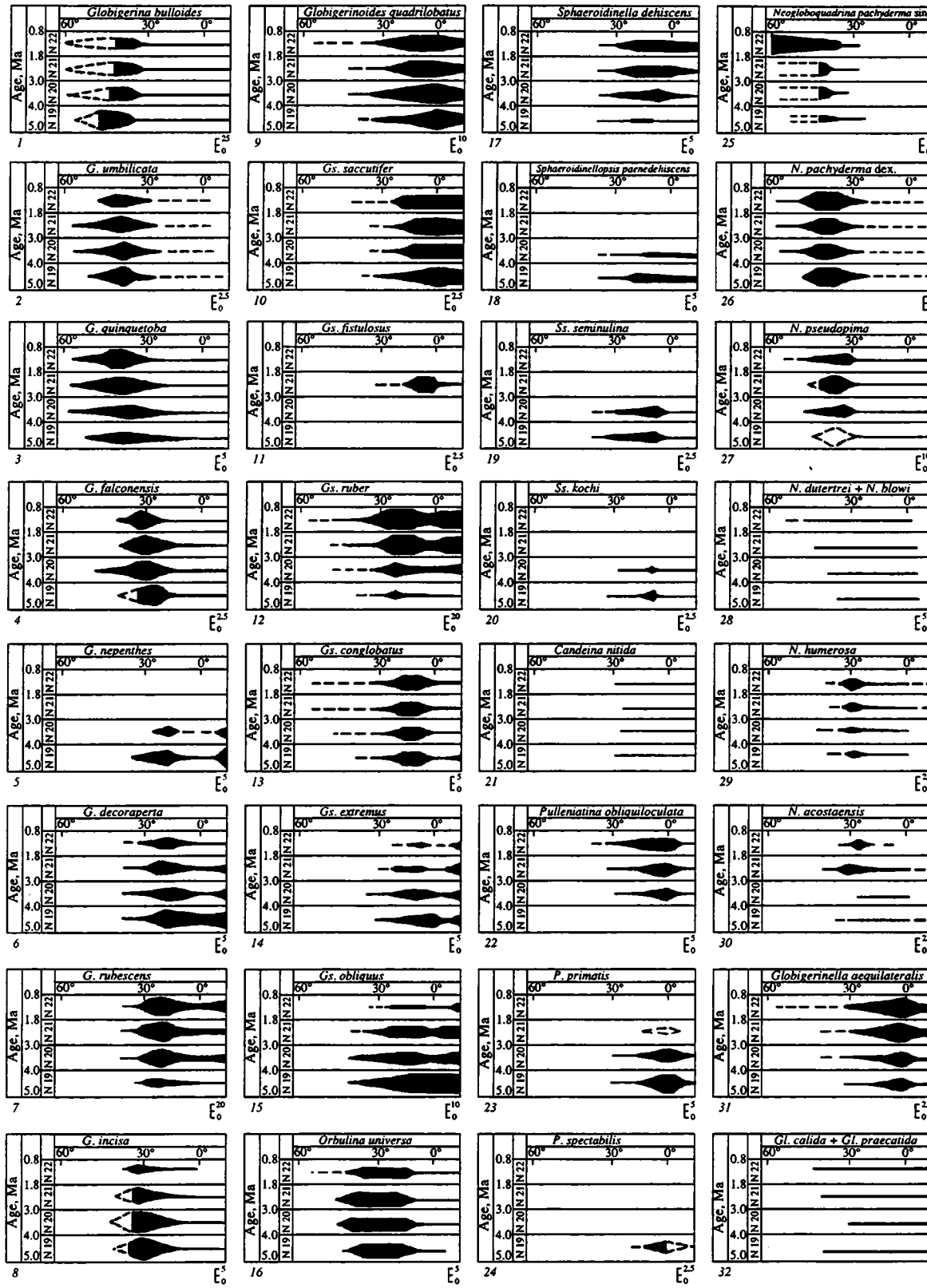
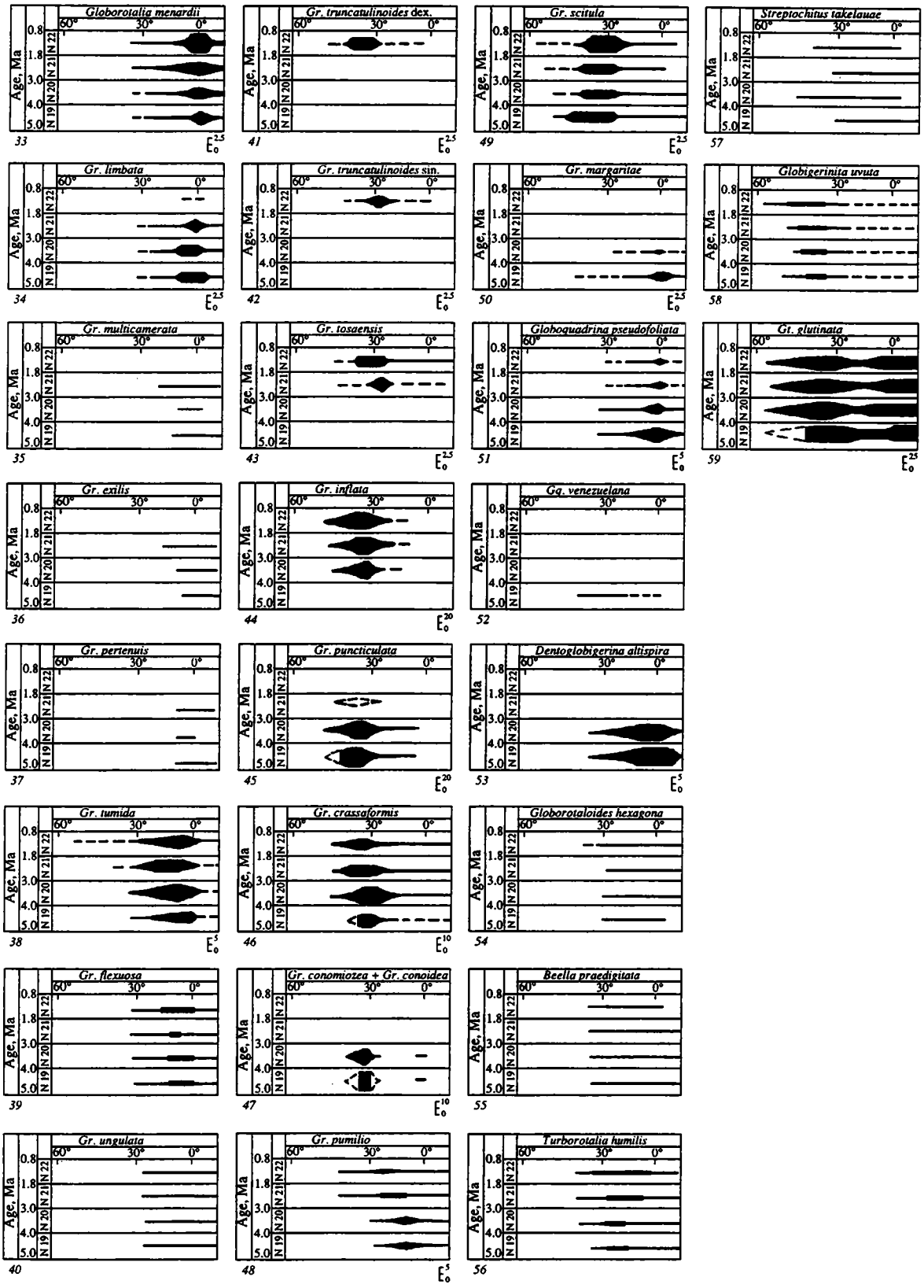


Fig. 3. Schemes of space–time changes in distribution areas of Pliocene–Quaternary planktonic foraminifer species in the North Pacific. Species are grouped according to their generic belonging with due account for species genetic relationships within the group. Scales under figures show the percentage.


Globorotalia menardii
Gr. truncatulinoides dex.
Gr. scitula
Streptochitus takelauae
Gr. limbata
Gr. truncatulinoides sin.
Gr. margaritae
Globigerinita uvuta
Gr. multicamerata
Gr. tosaensis
Globoquadrina pseudofoliata
Gt. glutinata
Gr. exilis
Gr. inflata
Gg. venezuelana
Gr. pertenuis
Gr. puncticulata
Dentoglobigerina altispira
Gr. tumida
Gr. crassaformis
Globorotaloides hexagona
Gr. flexuosa
Gr. conomiozea + Gr. conoidea
Beella praedigitata
Gr. ungulata
Gr. pumilio
Turborotalia humilis
Age, Ma
0.8
1.8
3.0
4.0
5.0
N 19
N 20
N 21
N 22
60°
30°
0°
33
41
49
57
34
42
50
58
35
43
51
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40
48
56

Fig. 3. (Contd.)

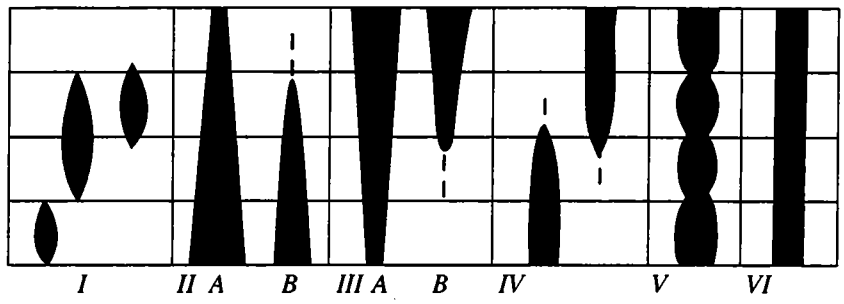

Fig. 4. Types of space–time changes of distribution areas of planktonic foraminifer species.
(I) short-lived; *(II) A*—reducing the distribution area, *B*—disappearing; *(III) A*—widening the distribution area, *B*—appearing and evolving; *(IV)* replacing each other; *(V)* slightly changing the distribution patterns; *(VI)* with stable distribution area.

Sphaeroidinella paenedehiscens–S. dehiscens (Fig. 3, *18* and *17*, respectively) are characterized by distribution areas which replace each other, similar abundance, and almost the same latitudinal dispersal and position of areas with their highest concentrations.

The above-described types of distribution areas (Fig. 4, *I* to *IV*) are characteristic of the planktonic foraminifer species, which are of significant value for the Pliocene–Lower Quaternary biostratigraphy. As it was shown, the distribution areas of the species, which were evolving during the considered time interval, changed gradually. Prior to final extinction, the species reduced their distribution area from the periphery (lowest abundances) toward the center (highest abundances). The newly appeared and evolving species widened their distribution areas from their centers toward peripheral parts; sometimes, areas of their highest abundances migrated in the same direction.

Thus, the same species could evolve in different latitudinal belts diachronously, and this should be kept in mind when estimating the age of sediments based on datum planes. The most reliable age values for the evolutionary appearance and extinction levels of particular species should be expected from the central part of their distribution areas. The longest duration of the species lifetime is also characteristic of their central part. Other age values can be related to local peculiarities of species dispersal. For instance, according to data by Dowsett (1989), the level of the evolutionary appearance of *Globorotalia truncatulinoides* in the southwestern Pacific is dated at 2.4 Ma, which is 0.5 Ma older than the value determined for this species in the Atlantic and North Pacific (Berggren *et al.*, 1984; and others). These observations show that the age of datum planes should be estimated separately in every climatic belt.

Two more types of space–time changes in distribution areas can be defined in addition to those described above (Fig. 4, *IV* and *V*).

(5) Distribution areas of species whose dispersal and abundance changed insignificantly throughout the Pliocene–Early Quaternary. These are species of high latitudes, such as *Globigerina bulloides, G. umbilicata, G. falconensis, Neogloboquadrina pachyderma* sinistral and dextral, and *N. pseudoopima* (Fig. 3, *1, 2, 4,* and *25* to *27*, respectively). As a rule, such species serve as indicators of certain climatic changes that result in the pulsation of distribution areas and species abundances. The ecological causes which were responsible for the dispersal of the species with distribution areas of this type are crucial for paleoceanographic interpretations.

(6) Distribution areas of species, which had stable dispersal throughout the considered time interval. These are *Globigerinoides quadrilobatus, G. sacculifer, G. conglobatus, Orbulina universa*, and *Globigerinita glutinata* (Fig. 3, *9, 10, 13, 15*, and *59*, respectively).

For some species with a concentration <5%, it is difficult to note any characteristic feature in their space–time distribution and to refer the distribution areas to any of the described types (Fig. 3, *3, 21, 32, 38, 39, 54*, and *56* to *58*).

CONCLUSION

Analysis of the stratigraphic distribution and dynamics of distribution areas of planktonic foraminifer species in space and time allowed the following conclusions to be drawn:

(1) In the Pliocene–Lower Quaternary sediments of the North Pacific, 65 species and varieties of planktonic foraminifers were identified and stratigraphic distribution of 60 species was established. Twenty five species of planktonic foraminifers became extinct, and ten species appeared during this time interval. Their diversity decreased from 53 species in the early Pliocene to 34 species in the Eopleistocene.

(2) According to the position of areas with the highest species concentration, we distinguished three groups of species displaying the highest abundances in lower, middle, and higher latitudes. These groups correspond to the three major biogeographic provinces of planktonic foraminifers in the global ocean: equatorial-tropical, subtropical, and temperate-subpolar ones.

(3) Abundant (>40%), common (40–15%), scanty (15–5%), and rare (<5%) species are distinguished in the studied assemblages of planktonic foraminifers.

(4) Changes in abundances and distribution areas of 60 species and varieties of planktonic foraminifers are outlined for time intervals corresponding to the N19, N20, N21, and N22 foraminiferal zones.

(5) Six types of space- and time-dependent changes in the species distribution areas were defined. These types reflect the peculiarities of planktonic foraminifer evolution in the North Pacific.

(6) Prior to extinction, species reduced their distribution areas and abundances from the periphery of the dispersal area toward its center, whereas the newly appeared and evolving species showed an opposite tendency.

(7) Genetically related species (ancestor–descendant) showed similar distribution patterns during their entire lifetime, and their dispersal areas sequentially replaced each other.

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Reviewer I.A. Basov

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Pliocene–Pleistocene Diatoms and Environments in the Northern Tartar Strait

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Abstract—Diatoms from a 34-m-thick member of Pliocene–Pleistocene sediments in the near-continental part of the Tartar Strait were studied. The beginning of the Pliocene was characterized by marine neritic environments indicating the stable influence of the Sea of Okhotsk waters. Warm-water environments were characteristic of the interval of 4.4–3.2 Ma, which is evidenced by subtropical diatom species. Their presence evidently reflects the periodically strong influence of warm Pacific waters and a weakening connection with the Sea of Okhotsk. The role of tectonics and climatic fluctuations was also considered.

Key words: *diatom algae, Pliocene–Pleistocene, Tartar Strait, geological events, biostratigraphy.*

Pliocene global events are of great interest, as they serve as a basis for the elaboration of prognostic models of environment evolution. Recently, diatom algae have been widely used in reconstructing the Pliocene climate dynamics. The diatom-based method of paleotemperature assessment (Td), which was suggested in the 1960s (Kanaya and Koizumi, 1966), is now the most popular quantitative method in paleoclimatic studies. For this purpose, the formula $Td = X_w/(X_w + X_c)$ is used, where X_w is the number of warm-water species and X_c is that of cold-water species. To obtain reliable temperature estimates, the number of identified diatom valves should not be less than 300 specimens.

The application of this method is most effective in the mid-latitude areas, where even insignificant changes in surficial water temperature are reflected in the species quantitative ratio (Barron, 1992b). In higher latitudes, cold-water species have dominated since the Pliocene, and changes in the ecological patterns of diatom associations are evident, first of all, at the qualitative level. They are mainly reflected in the increase of diatom diversity at the expense of tropical and subtropical immigrants, and in the appearance of warm-water silicoglagellate species and changes in the ratio between dominant groups of the arctic-boreal species (Oreshkina, 1990, 1993).

The data, which were obtained while studying diatom flora from a borehole drilled by R/V *Geolog Primor'ya* of AO Dal'morgeologiya on the continental shelf in the northern Tartar Strait, add new information on Pliocene–Pleistocene environments in the Far East. Borehole 2 (50°29′N; 140°41′E; water depth 66.3 m), drilled 8 km offshore near the Mednyi Cape, penetrated the 34-m-thick member of sands, silts and clays with shell debris and pebbles (Figs. 1 and 2). Twenty-five samples for diatom analysis, core description, and other relevant information were obtained from N.N. Kuz'mina, a researcher from the Laboratory of Quaternary Research at Moscow State University.

The diatom distribution patterns in the recent bottom sediments served as the basis for interpretations of

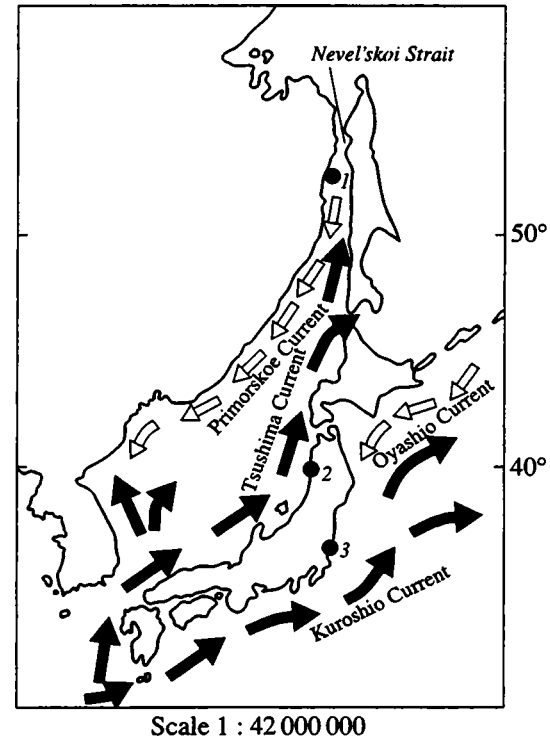

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Fig. 1. Scheme of modern circulation in the Japan Sea. (*1*) location of Borehole 2; (*2*) location of the Tatsunokuchi Formation section (Koizumi, 1973); (*3*) location of the Wakimoto Formation section in the Oga Peninsula (Saito and Burckle, 1977). White arrows—cold currents; black arrows—warm currents.

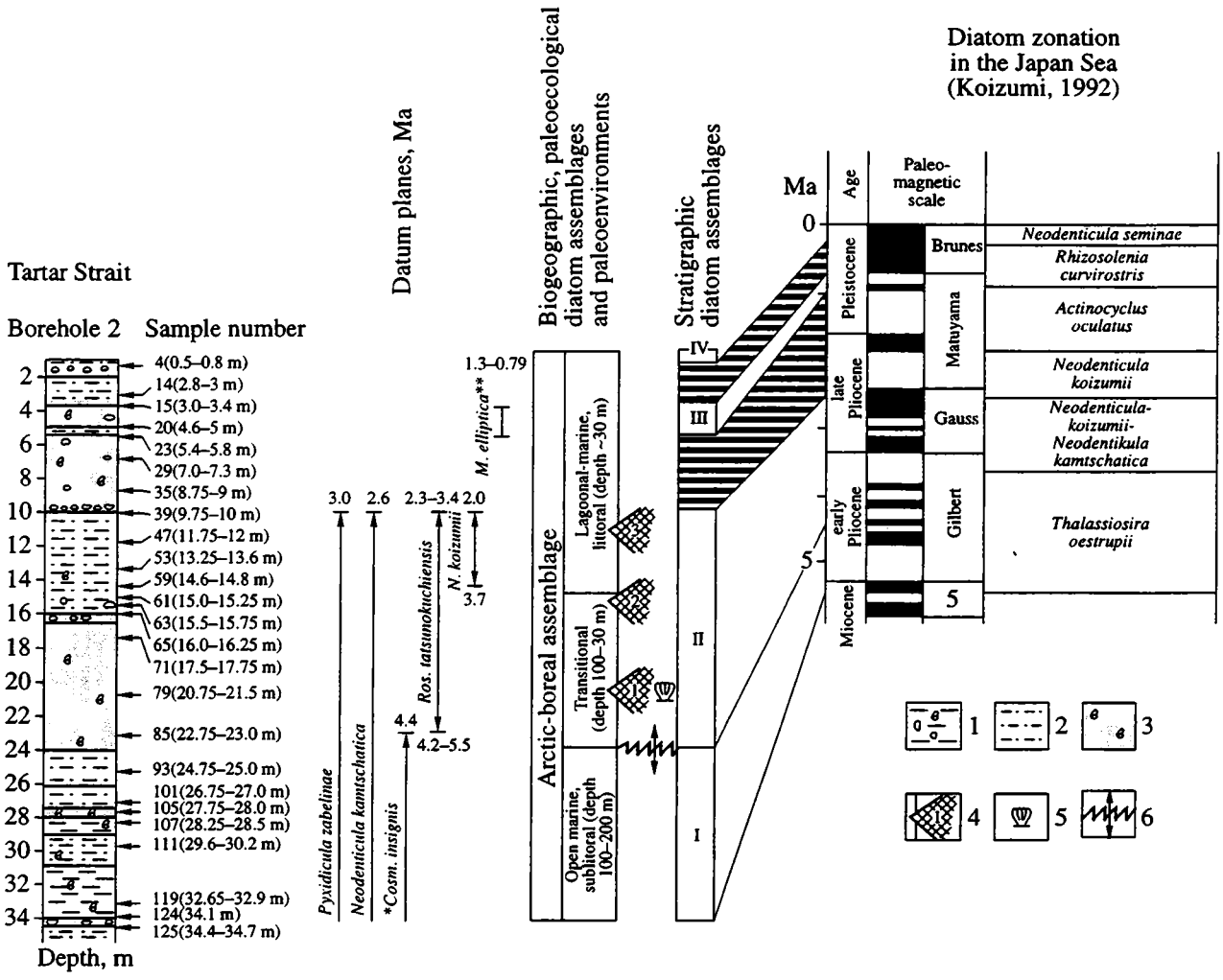


Fig. 2. Stratigraphic position of diatom assemblages in Borehole 2 and paleoenvironments.
(1) pelite with pebbles and shell debris; (2) silt; (3) sand; (4) invasions of warm water species; (5) *Fortipecten takahashii* occurrences; and (6) commencement of tectonic reorganization. Data on datum planes (Ma) after Barron, 1992 (marked by one asterisk), Burckle, 1979 (two asterisks), and Koizumi, 1992 (without mark).

the Pliocene paleoceanographic changes. Until recently, they have been poorly studied; at present, only general regularities of the recent diatom distribution are defined (Gail, 1936; Belyaeva, 1961; Jouse, 1962; Pushkar', 1979). The northern part of the Tartar Strait is located at the transition between the arctic-boreal and warm-water phytogeographic provinces, which results in the overlap of distribution areas of cold- and warm-water species and a mixed diatom composition. In the deeper oceanic assemblages, warm-water species dominate. Neritic assemblages of the shelf area are mainly represented by cold-water species. This is primarily related to the water circulation system in the Tartar Strait: the cold Primorskoe Current follows the continental coast, while the warm Tsushima Current, with its warm-water diatom assemblage, flows along the Japanese coast toward the southwestern extremity of the Sakhalin Island. However, north of 50°N, subtropical forms almost completely disappear from the diatom assemblages. The depths in the shallow-water Nevel'skoi Strait connecting the Sea of Okhotsk and Japan Sea vary from 1 to 27 m (*Atlas* ..., 1967).

The recent diatom assemblages also feature the presence of older fossil species which can be explained by their reworking from the Tertiary deposits exposed at the sea bottom, and also, from diatomites and similar rocks widely represented in onshore outcrops in Japan and Sakhalin.

The inner shelf facies commonly contain abundant *Paralia sulcata* (Ehr.) Cl., which inhabits the sandy shoals, and sponge spicules, whose mass accumulations are confined to the depths of 25 to 100 m. It is believed that the mass development of benthic diatoms is limited by the 30-m isobath.

In the present paper, the stratigraphic subdivision and dating of the sediments were performed by means

of datum planes, which are now well known (Barron, 1992a; Koizumi, 1992). The application of classical zonal stratigraphy in this inner sea basin, which is located at the joint of cold and warm provinces, encounters certain difficulties. These are caused by a diachrony in respective levels of the first and last occurrences of stratigraphic markers, and abrupt transitions between warm- and cold-water assemblages, which resulted from climatic fluctuations and tectonic-related current shifts. When studying shelf facies stratigraphy, one cannot avoid the negative consequences of sediment reworking. As a rule, the stratigraphically important assemblages of high diversity with zonal oceanic species are commonly enclosed in very fine-grained rocks, which correspond to the maximum transgression. The neritic assemblages are correlated with the zonal scale by means of the interpolation method while taking into consideration sedimentation rates and cyclicity.

In Borehole 2, diatoms were present in all the samples, but their abundance and diversity show significant variations in different samples. The stratigraphically important diatom assemblages in some intervals allow their age to be determined with sufficient accuracy and ecological patterns to be defined.

The interval of 34.9–25 m, mainly composed of silts and clays, is referred to the early Pliocene (5.1–4.4 Ma). It contains the neritic assemblage with dominant boreal and arctic-boreal species: *Pyxidicula* (Thalassiosira) *zabelinae* (Jouse) Makar. et Mois., *Thalassiosira jouseae* Akiba, *T. gravida* Clevs, *Actinocyclus curvatulus* Janisch, *Neodenticula kamtschatica* (Zabelina) Akiba and Yanagisawa, *Cosmiodiscus insignis* Jouse, *Rhizosolenia setigera* Bright., *Coscinodiscus marginatus* (Ehr.), *C. oculus-iridis* Ehr., and *Nitzschia cylindrus* (Grun.) Hasle. The presence of *Thalassiosira oestrupii* (Ost.) Proshk.-Lavr. is the most important stratigraphic marker. The fragmented valves of the freshwater species *Pinnularia viridis* (Nitz.) Ehr., whose presence evidently reflects the influence of the Amur River waters, are a constant constituent of the assemblage. It is pertinent to note that the first appearance of the *Pinnularia* genus representatives occurred in the pre-Pliocene time. Also present are single specimens of the late Miocene thick-walled species most resistant to valve destruction: *Porosira punctata* (Jouse) Makar. and *Pyxidicula inermis* (Jouse) Streln. et Nikolaev. Some levels are marked by an increase in the abundance of sponge spicules. This is associated with the simultaneous decrease in diversity and increase in abundance of tychopelagic and benthic diatom species (Samples 125, 105, and 93). The latter events commonly result from the shoaling of the basin. In Sample 105, the increase in the share of thermophilic forms *Thalassiosira lineata* Jouse and *T. oestrupii* (Ost.) Proshk.-Lavr. is recorded.

It is interesting that the deepest and generally cold-water neritic diatom assemblage associates with the warmest (for Borehole 2) spore and pollen complex (data by G.I. Shumova). The discrepancy in data on marine and terrestrial biotas is most likely explained by the strong water exchange between the considered basin and the Sea of Okhotsk, i.e., by the opening of the Nevel'skoi Strait as a result of tectonic reorganization. It should be noted that, according to Borehole 2 lithology, the considered interval corresponds to the period of maximum depths in the basin.

The interval of 23–15 m (fine-grained sands) is also assigned to the early Pliocene. The position of the Pliocene lower boundary is based on the last occurrence level of *Cosmiodiscus insignis*, the species which is typical of the North Pacific diatom flora. According to the available data, this event is synchronous in the mid- and high-latitude areas and occurred 4.4 Ma ago. The sediments of this interval incorporate the assemblage, which contains, along with the less abundant above-listed species, the first appearance of the abundant diatom species *Thalassiosira lineata* Jouse, *T. delicata* (Barron) Akiba, *T. oestrupii* (Ostenf.) Hasle, *T. antiqua* (Grun.) Cl., *T. eccentrica* (Ehr.) Cl., *Rhizosolenia styliformis* Briht., *Rossiella tatsunokuchiensis* (Koizumi) Gersonde et Schrader, *Actinocyclus ehrenbergii* Ralfs, and *Thalassionema nitzschioides* Grun., which are more typical of the mid-latitude areas. The sporadic occurrence of *Detonula confervaceae* (Cl.) Gran., one of the representatives of the recent endemic diatoms, is also recorded.

From the level of 15.75 m upward, an ecological transformation of diatom assemblages is evident. The marine benthic species of the *Raphoneis* group represented by *Raphoneis* aff. *ishaboensis* (Grun.) Mertz, *R. margaritata* Andrews, and *Delphineis surirella* Ehr, whose abundance gradually increases upsection, becomes the most noticeable component of the diatom assemblage. The predominance of benthic species suggests an abrupt decrease in deposition depths and lagoonal-marine environments, which probably resulted from the Nevel'skoi Strait closing.

The remarkable feature of Sample 79 (interval of 20.75–21.50 m) is the highest abundance in of the form most similar to *Coscinodiscus gorbunovii* first described by V.S. Sheshukova-Poretskaya (Sheshukova-Poretskaya and Moiseeva, 1964). Later, this species was assigned to the new genus *Pontodiscus* [*P. gorbunovii* (Sheshuk.) Moiss. et Sheshuk.] (Temniskova-Topalova *et al.*, 1982) and then to the *Undatodiscus* genus, and was described as *U. sheshukovae* Lupik. (Lupikina, 1982). Note that in the considered samples, the specimens assigned to this form bear distinct areole fascicles, distinguishing them somewhat from the latter. *Coscinodiscus gorbunovii*, which is thought to be a freshwater species, is widely distributed in the Suifun Formation of the Primor'e region and Erman Formation of western Kamchatka. Both these formations, according to palynological data, are referred to the Pliocene and reflect the temperate warm deposition environments. One can suggest that the constant presence of

this species testifies to the connection of the considered basin with the large continental freshwater basins of the Primor'e region that appeared either during major transgressions or due tectonic reorganizations.

In addition, this interval is marked by the appearance of some thermophilic diatom species, such as *Azpeitia tabularis* (Grun.) Fryxell et Sims, and by the increase in abundance of *Thalassiosira antiqua* Grun. (Sample 63).

The assemblage, which is almost identical to that recorded in the interval of 23–15 m, was described (Koizumi, 1973) from the Tatsunokuchi Formation (the Miyagi Prefecture, Japan) on the Pacific coast of Honshu Island (38°N). This about 50-m-thick formation contains, along with the aforementioned diatoms, the bivalves *Anadara tatunokutiensis, Dosinia tatunokutiensis*, and *Fortipecten takahashii* assigned to the lower Pliocene. It should be noted that, according to E.A. Taldenkova (personal communication), the first occurrence of *Fortipecten takahashii* in Borehole 2 corresponds to the considered interval.

Invasions of thermophilic forms into the northern part of the Tartar Strait 4.4–3.0 Ma ago and the similarity of diatom assemblages from the studied area and synchronous assemblages of the Pacific coast of Honshu Island most likely suggest water circulation patterns essentially different from the recent ones. It should be noted that the materials of ODP Legs 127/128 in the Japan Sea (Koizumi, 1992) testify to the weakening influx of warm Pacific waters of the Tsushima Current into the Japan Sea throughout the Pliocene. This is evident from the absence of thermophilic forms from the marine biota in the southern Japan Sea. It is also remarkable that the climatic optimum in terrestrial flora evolution occurred all over the Japan Sea region at that time. In the Japan Sea, the warmest palynocomplexes with the lowest share of spores and pollen of coniferous and rich in specimens of warm-temperate broad-leaved forests are dated at approximately 4 Ma (Heusser, 1992). This age estimate is consistent with the data on diatoms.

The available data show that marine deposits containing similar diatom assemblages and assigned to the lower Pliocene are widely distributed in the southern Kuril Islands, southern part of Sakhalin and on the southwestern Sakhalin shelf (Jouse, 1962; Margulis *et al.*, 1980; personal data). This might be a result of both eustatic transgression and regional tectonic events.

Upsection (interval of 14.8–13.8 m), the diatom assemblage again acquires cold-water features. The diatom abundance decreases and only cold-water species are present.

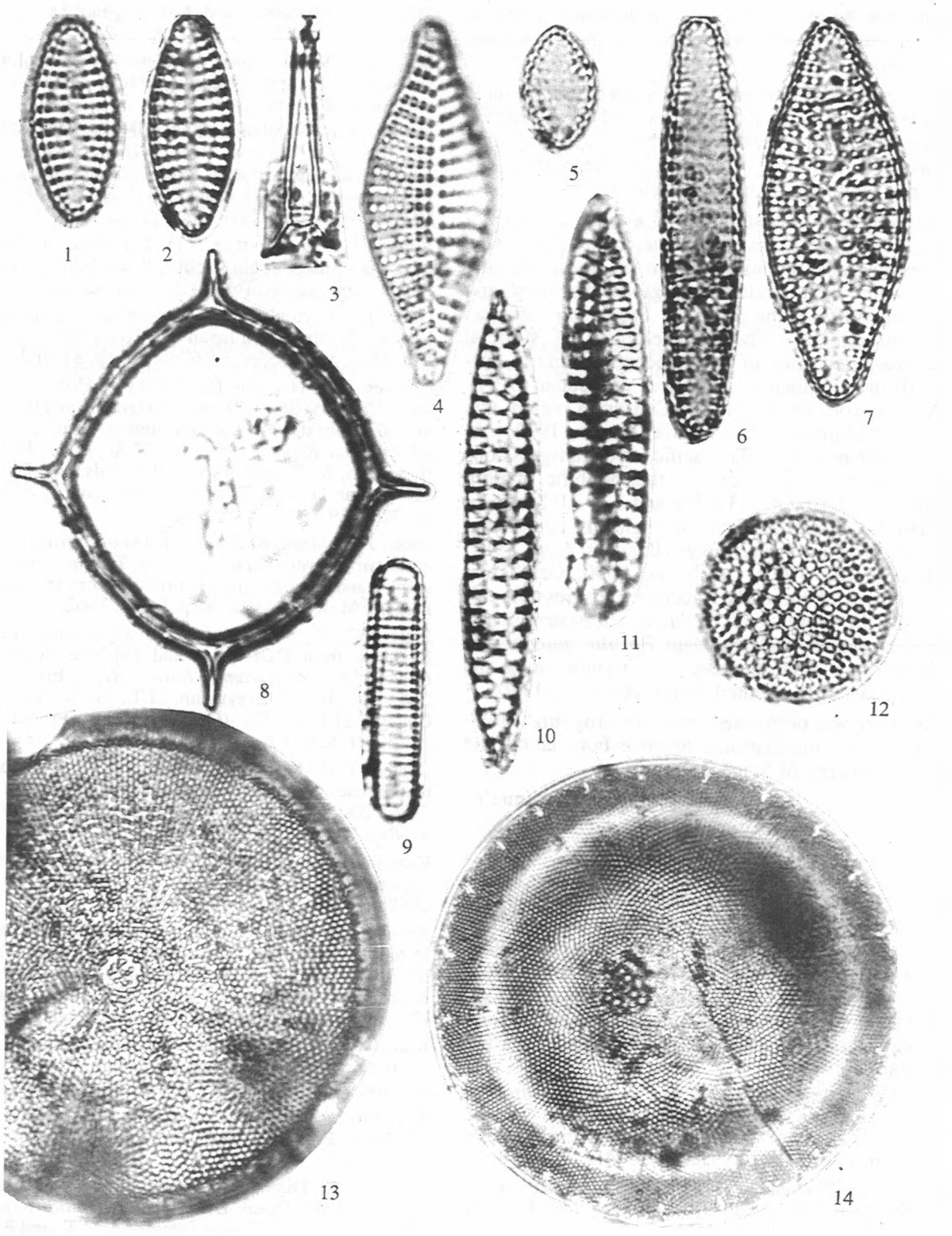
At the level of 12 m, the warm-water assemblage appears once again, being dominated by the south-boreal and subtropical species *Rossiella tatsunokuchiensis, Bacteriastrum hyalinum* Lauder, and *Actinocyclus ehrenbergii*. Also present are single valves of *Neodenticula koizumii* Akiba and Yanagisawa, whose first occurrence at the periphery of the arctic-boreal province is recorded at about 3 Ma. In view of the fact that the last occurrence of *Rossiella tatsunokuchiensis* in the northern Japan Sea corresponds to about 3.2 Ma (Koizumi, 1992), the age of the considered assemblage is interpreted to be approximately 3 Ma.

The sediments in the interval of 10–5.8 m are characterized by a diatom assemblage of no stratigraphic significance. The presence of single specimens of *Neodenticula koizumii* is the only stratigraphic marker that allows the age of the sediments to be determined. As it was noted earlier, this form appears at the same latitudes of the Pacific, approximately 3 Ma ago.

The overlying sands and silts (interval of 5.8–2.8 m) incorporate the assemblage with *Mesocena elliptica* (Ehr.) Defl., referred to the lower Pleistocene. In addition to this silicoflagellate species, the sediments contain a mixed diatom assemblage composed of warm- and cold-water species: *Detonula confervaceal, Raphoneis* aff. *ishaboensis, R. surirella, Thalassionema nitzschioides, Thalassiosira gravida, T. jouseae, Porosia glacialis* Jorg., *Actinicyclus oculatus* Jouse, *A. ochotensis* Jouse, *A. ehrenbergii*, and *Pseudopyxilla rossica* (Pant.) Forti. Also present are freshwater benthic forms. *Pyxidicula zabelinae* is represented only by valve fragments that suggest reworking. An age estimate of this assemblage is based on various data (Kazarina, 1975; Saito and Burckle, 1977; Burckle and Opdyke, 1977; Burckle, 1979) indicating that the last occurrence of *Mesocena elliptica* in the entire Central Pacific is a diachronous event, which is recorded within the interval from the base of the Jaramillo magnetic epoch to the base of the Brunhes Chron, i.e., within the 1.3–0.79 Ma time interval. The assemblage with *Mesocena elliptica* was also reported from the Wakimoto Formation in the Oga Peninsula, the western coast of Honshu Island (Saito and Burckle, 1977).

The age of the highest diatom assemblage (interval of 0.5–0.8 m; mainly sands) is also interpreted in different ways. It lacks extinct species, with the exception of undoubtedly reworked thick-shelled Miocene forms (*Actinocyclus ingens* Rattr. and *Ikebea tenuis*). The assemblage is dominated by *Paralia sulcata*. On the one hand, the latter can be regarded as a typical Holocene species, because its mass occurrence is in accord with the recent environments at the drill site. On the other hand, it was established that in the Primor'e region, similar assemblages dominated by the tychopelagic species and reworked Tertiary forms correspond to the Sartan glacial epoch. Therefore, additional information is needed to unambiguously solve this problem.

Thus, the biostratigraphic data on diatoms show that the considered borehole penetrated the Pliocene–Pleistocene sediments. The diatom assemblages, which are important for paleoceanographic interpretations, are present only in the lower Pliocene interval. Three levels where the south-boreal and subtropical species predominate are defined: at 15–23 m, 15 m, and 11.35–12 m.



Reference diatom species from Pliocene–Quaternary deposits of the northern Tartar Strait.

(1, 2, 7) *Delphineis surirella* (Ehr.) Andrews: (1, 2) Sample 85, ×2000; (7) Sample 47, ×2000. (3) *Rhizosolenia styliformis* Bright., Sample 63, ×2000. (4) *Raphoneis margaritata* Andrews, Sample 85b, ×1500. (5, 6) *Raphoneis* aff. *ishaboensis* (Grun.) Mertz: (5) Sample 85, ×2000; (6) Sample 63, ×2000. (7) *Mesocena elliptica* Ehr., Sample 15, ×1000. (8) *Nitzschia cylindrus* (Grun.) Hasle, Sample 85b, ×2000. (10, 11) *Rossiella tutsunokuchiensis* (Koizumi) Gersonde et Schrader: (10) Sample 47, ×2000; (11) Sample 85, ×2000. (12) *Thalassiosira oestrupii* (Ost.) Proshk.-Lavr., Sample 47, ×2000. (13) *Actinocyclus ehrenbergii* Ralfs, Sample 47, ×1000. (14) *Coscinodiscus gorbunovii* Seshuk., Sample 55, ×2000.

In addition, benthic diatoms are a dominant group at 15.75 m, which, evidently, indicates the decreasing accumulation depths.

The changes in diatom assemblages reflect unstable tectonic environments in the considered area, which is in agreement with recent idea on the intense formation of thrusts and faults in the Japan Sea during the Pliocene (Melankholina, 1987; Ingle, 1992).

The influence of eustatic transgressions also cannot be excluded. They may contribute to the considered changes in the circulation system of Japan Sea and Pacific waters that resulted from global warmings and northward shifts of the Subarctic Front. For instance, the warm interval, when temperatures of surficial waters exceeded those of the Holocene maximum, corresponds in the Subarctic Front area near Japan to 3.1–3.0 Ma (Barron, 1992b). In the California region, this event was recorded at 3.7–3.0 Ma (Barron, 1992c). In the Far East region of the Pacific, two transgressions with respective invasions of thermophilic diatom assemblages during 4.5–3.8 Ma and 3.6–3.0 Ma were revealed by materials from Karaginskii Island and Deep Sea Drilling Project holes (Basilyan *et al.*, 1991, Basilyan *et al.*, 1993; Oreshkina, 1990; Oreshkina, 1993). Traces of the early Pliocene transgression evidenced by the presence of *Thalassiosira oestrupii*, and *Rossiella* species, and abundant *Paralia sulcata* and *Raphoneis* are also recorded in Virginia along the Atlantic coast of the United States (Andrews, 1980).

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The Chekalin (Likhvin) Reference Section in the Russian Plain and Its Significance for Pleistocene Stratigraphy and Correlation

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Abstract—The completeness of the comprehensively studied stratigraphic column comprising the parastratotype of the Belovezh'e Horizon and the stratotypes of the Oka and Likhvin horizons illustrates the great significance of the Likhvin reference section for correlation. A detailed study revealed the age relationships and facial transitions between heterogenous deposits, and allowed us to define the exact positions of the Oka moraine and Likhvin stratotype proper. The periglacial environments were proven to exist prior to the Belovezh'e interglacial stage. This served as the basis for recognizing the Don Horizon in the section. A better understanding of the climatostratigraphic peculiarities of the Belovezh'e palynoflora was achieved. Climatic rhythms were reconstructed for the entire Oka–Dnieper stratigraphic interval. The main objectives were to elucidate the representability and completeness of the Likhvin stratotype, to refine its climatostratigraphic subdivision, and to reveal characteristic features of other interglacial horizons in the section. The climatostratigraphic structure of the Likhvin–Dnieper interval is specified, and new units of the Likhvin Superhorizon are itemized to correct the regional stratigraphic scheme.

Key words: *Stratigraphy, correlation, paleogeographic reconstructions, lithology, mineralogy, palynology, paleovegetation, paleoclimate, glaciations, interglacials, stratotype.*

The reference stratotype section near the town of Chekalin (Likhvin) is remarkable in its Quaternary geology, because the unique Pleistocene stratigraphic column comprises here a full succession of lower and middle Pleistocene units. It is not by chance that this section has been long studied (Bogolyubov, 1904; Sukachev, 1907; Moskvitin, 1931, 1967; Ushko, 1959; Grichuk, 1961; Ananova and Kul'tina, 1965; Sudakova, 1973; Bolikhovskaya, 1974; *Razrezy* ..., 1977; Breslav *et al.*, 1981, and others).

The 40-m-thick section at the left bank of the Oka River exposes glacial, lacustrine, alluvial and loess deposits, which alternate with fossil soils and cryogenic horizons and show regular facial transitions along the strike (Fig. 1). Paleogeographic data derived from the Likhvin section are stratigraphically advantageous, because the clear chronological relationships between the moraines and loess-soil complex allows us to correlate this reference section, located in the glaciation area of the Russian Plain, with sections of southern periglacial loess provinces. The significance of the whole section comprising the well-known reference stratotype stimulates the current interest in its comprehensive examination. At present, the necessity to obtain additional data stems from controversial views on the quantity and character of paleoclimatic rhythms recorded in the complex Oka–Dnieper interval, from different opinions on the position and completeness of the Likhvin stratotype in terms of the "Great Likhvin" event, and from possibility to reveal landscape and climatic changes during the Don–Oka interval of the Early Pleistocene.

New palynological and lithological investigations enable us to elucidate these problems and, particularly, to reconstruct in detail the middle Pleistocene climatic rhythms and specify the stratigraphic position of the Likhvin stratotypical oxbow lense in the lower-middle Pleistocene sequence and its relationship with the Oka moraine. Ten sedimentary complexes, different in age and subdivided into 33 beds of diverse lithologic composition, genesis, and sedimentation settings, were recognized in the section and thoroughly studied.

The Don Horizon

The pre-Oka sequence recovered by the deep drilling is represented by alternating sandy and clayey sediments of lake and basin origin (Beds 2–7) (Figs. 1–3).

The accumulation of sandy-pebbly morainelike deposits (Bed 1) was related to the development of the ancient Oka depression in the early Pleistocene. These deposits are overlain by 6-m-thick clayey lacustrine sediments (Bed 2), which grade upward into the morainelike deposits (Bed 3). The fluvio-glacial genesis of clays in Bed 2 is evidenced by the predominance (60% and more) of pollen and spores of cryophytes (*Betula nana, B. fruticosa, Alnaster fruticosus, Dryas octapetala, Selaginella selaginoides, Botrychium boreale*, and others), noticeable amount of pollen of hydrophytes (*Myriophyllum, Potamogeton*, and others), and abun-

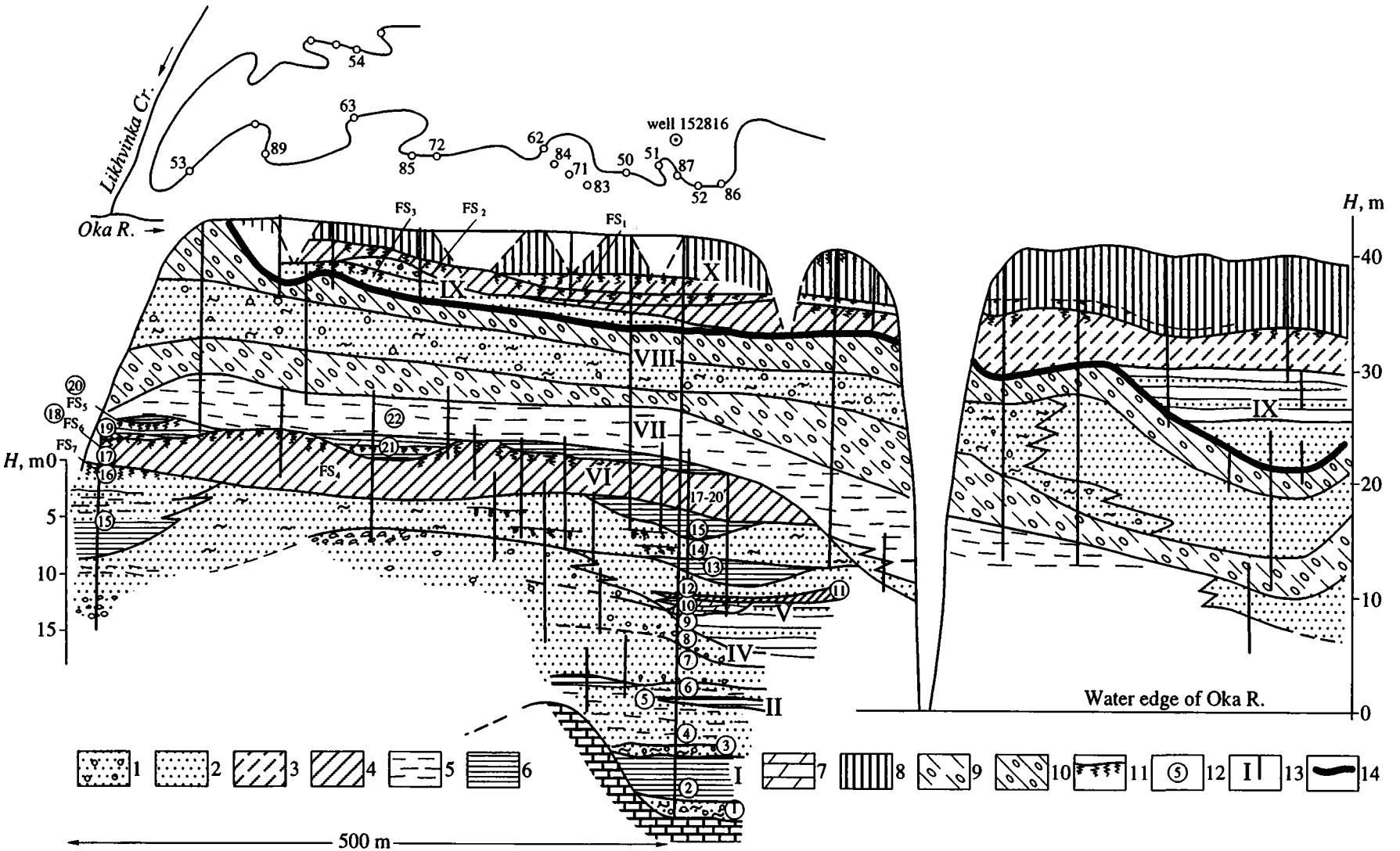


Fig. 1. The Pleistocene section near the town of Chekalin (Likhvin) (1) sand with pebble; (2) sand; (3) loamy sand; (4) loam; (5) silt; (6) clay; (7) foliated marl; (8) blanket loesslike loam; (9) boulder loamy sand (moraine); (10) boulder loam (moraine); (11) fossil soil; (12) bed number; (13) cuttings and wells; (14) a top of the Dnieper moraine.

dant pre-Cenozoic palynomorphs (in this case, spores of the Carboniferous *Waltzispora, Diatomozonotrilites, Trolobozonotrolites*, and others), which are characteristic of the glacial palynospectra reported from the Russian Plain (Fig. 2). At that time, the Oka upper valley was occupied by the periglacial forest-tundra vegetation dominated by shrub communities of dwarf birch, alder groves, *Betula fruticosa*, and possibly willows in association with green moss, sedge fens, and meadow to meadow-bog coenoses of grasses, sedges, herbs, grape-ferns, and others. The light pine–spruce forests with admixtures of *Pinus sibirica* and *Betula pubescens* sporadically occurred in the most favorable habitats. In addition to the specified climatic phytocoenoses and facies, the significant total thickness of Beds 1–3 also suggests their relation to the long-term Don glaciation.

The Belovezh'e Horizon

The overlying 10-m-thick lacustrine sequence consists of horizontally bedded sands with clayey laminae. It reflects two stages of sedimentation: the first stage corresponds to Beds 4–6 and the second one, to Bed 7, each unit displaying a grain-size trend decreases upward the section. The entire sequence bears a single staurolite–ilmenite mineralogical association sharply differing from that of the overlying sediments because of its insignificant content of exotic hornblende and garnet. This shows that local sources of mineral transportation into lake basins were most significant.

The stratigraphic position of Beds 4–7 and the prevalent pollen spectra of interglacial forests (Fig. 2) suggest that the respective deposits were formed during the optimum climatic phases of the Belovezh'e Interglacial separated by an endothermal.[1] The initial and terminal phases of the interglacial rhythm are not represented in the section. The sands of Beds 4, 6, and 7 yielded pollen spectra, indicating the domination of spruce and thermophilic broad-leaved plants associated with thermophilic exotics. Because the gray clays of Bed 5 contain the interstadial type macroflora (Velichkevich, 1982) and pollen spectrum of the spruce forest type, this bed can be related to the endothermal. The Carboniferous spores (up to 25%), as well as spores and pollen of cryophytes from its upper part, do not exclude the periglacial origin of the bed, which is likely related to the Don accumulation area as a source of redeposited material.

An assemblage of characteristic components of the Belovezh'e palynoflora from the Likhvin reference section can serve as an important stratigraphic marker. The assemblage includes *Cedrus* (autochtonous element of long-distance transportation), *Tsuga canadensis* L., *Picea* s. *Omorica*, *P.* s. *Eupicea*, *Abies* sp., *Larix* sp., *Pinus* s. *Cembra*, *P.* s. *Strobus*, *Rhus* sp., *Carpinis betulus* L., *C. orientalis* Mill., *Fagus sylvatica* L., *Quercus robur* L., *Q. pubescens* Willd., *Tilia cordata* Mill., *T. platyphyllos* Scop., *T. amurensis* Rupr., *Ilex aquifolium* L., *Ulmus laevis* Pall., *U. glabra* Mill., *U. campestris* L., *Osmunda cinnamomea* L., *O. claytoniana* L., *O. regalis* L., *Woodsia fragilis* (Trev.) Moore., *W. manchuriensis* Hook., and *Salvinia* sp.

The palynozones Bv1–Bv4 distinguished in the spore-pollen diagram (Figs. 2 and 3) correspond to a number of finer climatostratigraphic subdivisions of the Belovezh'e Horizon. Their characteristic spectra allow us to reconstruct the relevant phases and subphases of vegetation development in this period:

Bv1—zone of spruce, pine, hornbeam, oak, lime, elm, and alder (phase of polydominant mixed forests):

1a—subphase of pine–spruce and broad-leaved (elm-lime-hornbeam) forests with alder groves;

1b—subphase of broad-leaved (elm–oak–hornbeam) forests, alder groves, and pine–spruce forests (the first climatic optimum);

Bv2—zone of spruce (phase of spruce forests of the endothermal cooling);

Bv3—zone of spruce, pine, birch, oak, elm, hornbeam, and alder (phase of pine–birch and spruce–hornbeam–elm–oak forests and alder groves);

Bv4—zone of spruce, broad-leaved trees, and alder (phase of mixed forests):

4a—subphase of spruce and broad-leaved (elm–oak–hornbeam) forests with an admixture of lime, pine, and birch; hazelnut undergrowth and alder groves;

4b—subphase of spruce–elm–oak–hornbeam forests with an admixture of hemlock, cedar, fir, larch, beech, holly, pines and birches (the second climatic optimum); the grass cover was dominated by ferns of the Polypodiaceae and Osmundaceae families (*Osmunda cinnamomea, O. claytoniana*, and *O. regalis*);

4c—subphase of spruce–hornbeam–elm–lime and birch–pine forests and alder groves.

The Oka Horizon

The Oka (Bryan'kovo) moraine proper, up to 5 m thick, is exposed in neighboring outcrops at the Likhvinka Creek and Bryan'kovo Village. It occurs in the valley margins near the current water line, hypsometrically higher than the alluvial-lacustrine deposits. The moraine is represented by very compact greenish-brown loam enclosing pebble and gravel of predominantly sedimentary deposits. Exotic clasts of Scandinavian origin (granites, metamorphic and basic rocks) constitute about 5%. The matrix bears the staurolite–ilmenite mineral assemblage with an admixture of hornblende (15–20%). According to these features, the Oka moraine is quite distinct from the Dnieper one. The distinguished paleomagnetic "Bryan'kovo" episode (Sudakova and Bol'shakov, 1977) is also important for correlation, as it proves the correlation between the

[1] Bolikhovskaya (1991) proposed this term for the intervals of the relatively cool and dry climate between the thermal interglacial maxima.

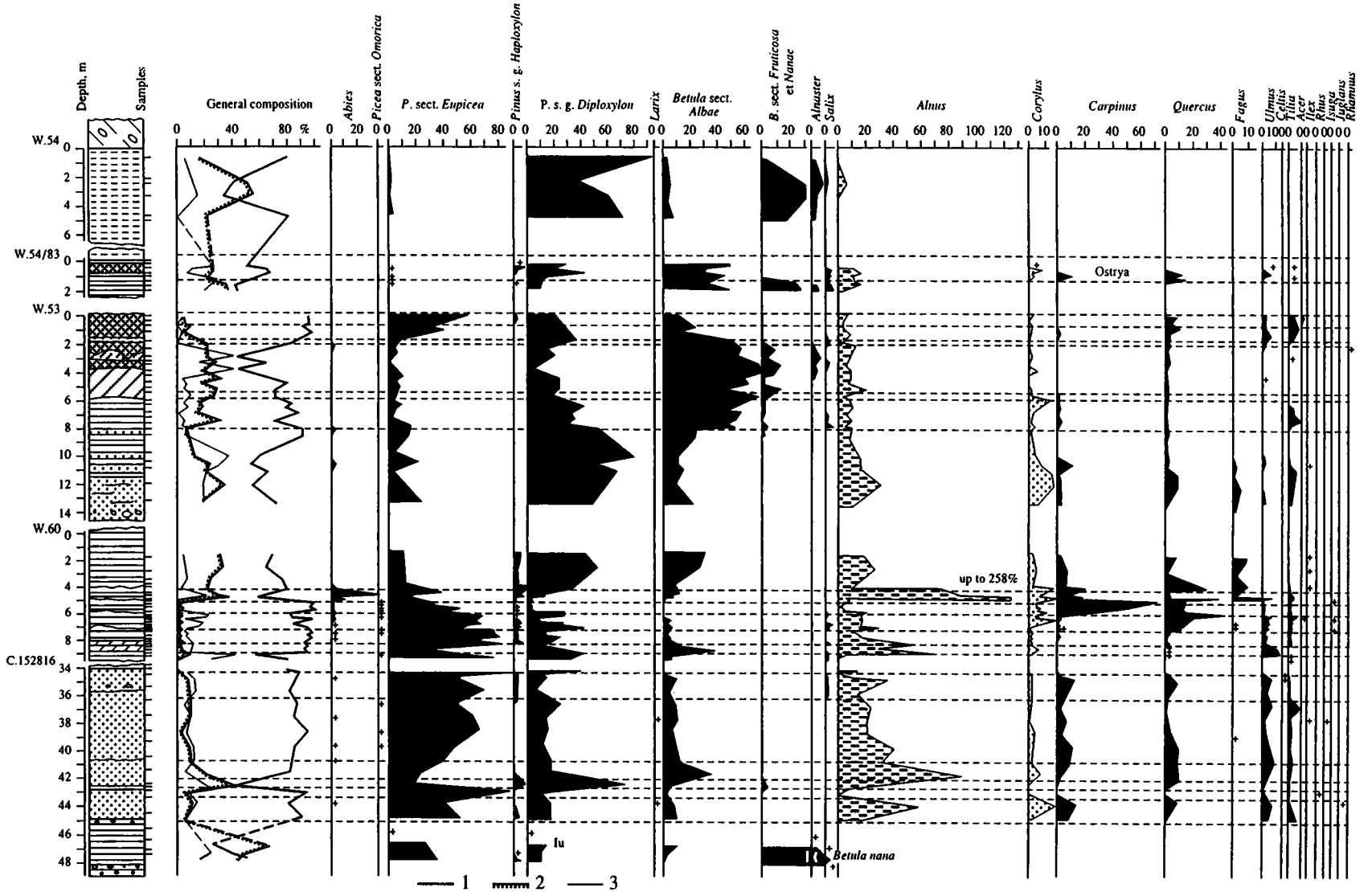
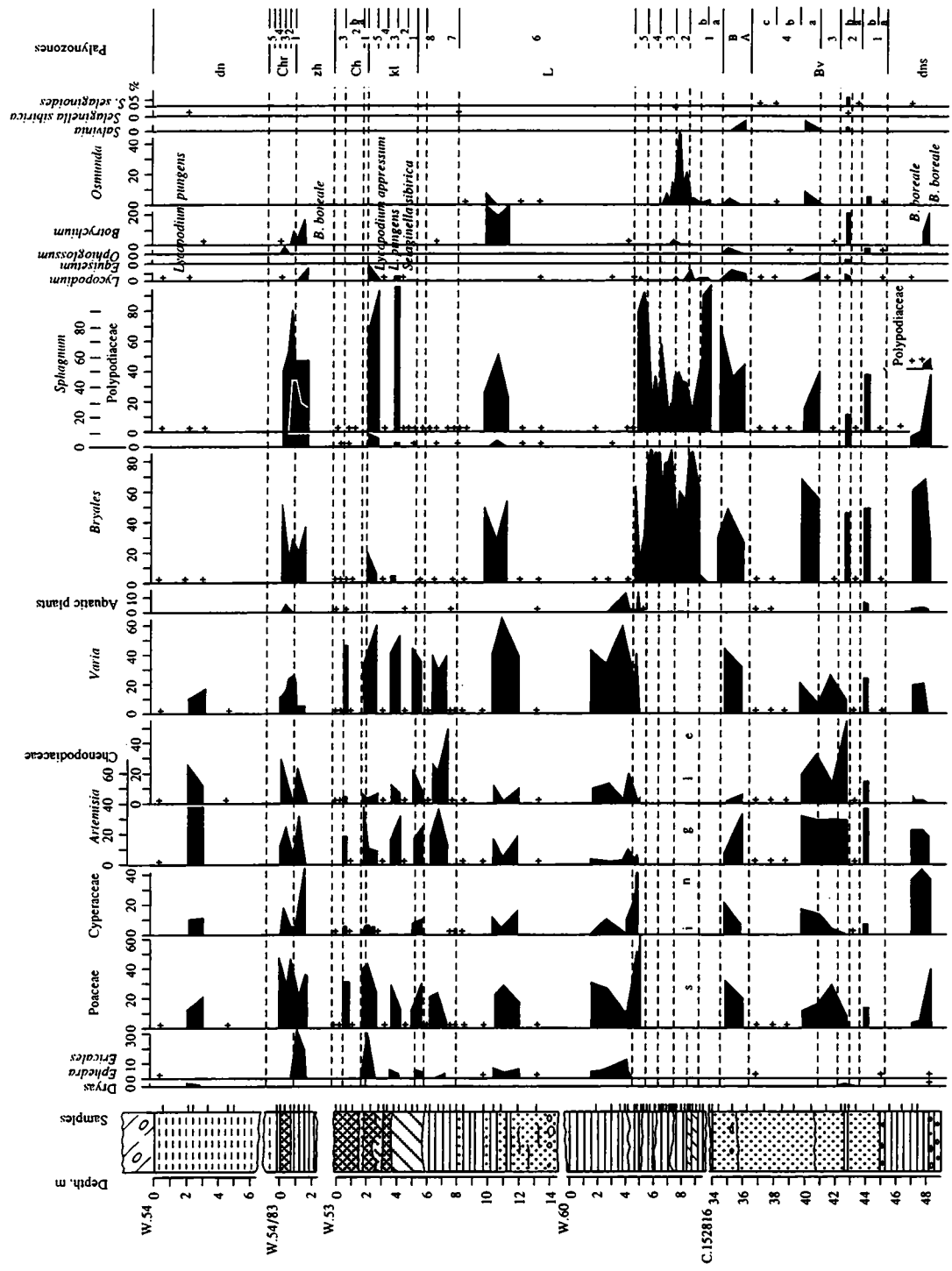


Fig. 2. Spore-pollen diagram of the undermoraine deposits of the Likhvin section: (1) pollen of tree species; (2) pollen of herbs and undershrubs; (3) spores. For column symbols see Fig. 1.


Palynozones
dn
Chr
zh
Ch
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L
Bv
dns
Selaginella sibirica
S. selaginoides
Osmunda
Lycopodium pungens
B. boreale
Borychium
Lycopodium
Equisetum
Ophioglossum
Lycopodium appressum
L. pungens
Selaginella sibirica
B. borreale
Sphagnum
Polypodiaceae
Bryales
Aquatic plants
Varia
Chenopodiaceae
Artemisia
Cyperaceae
Poaceae
Dryas
Ephedra
Ericales
Samples
Depth, m
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W.54/83
W.53
W.60
C.152816

Fig. 2. (Contd.)

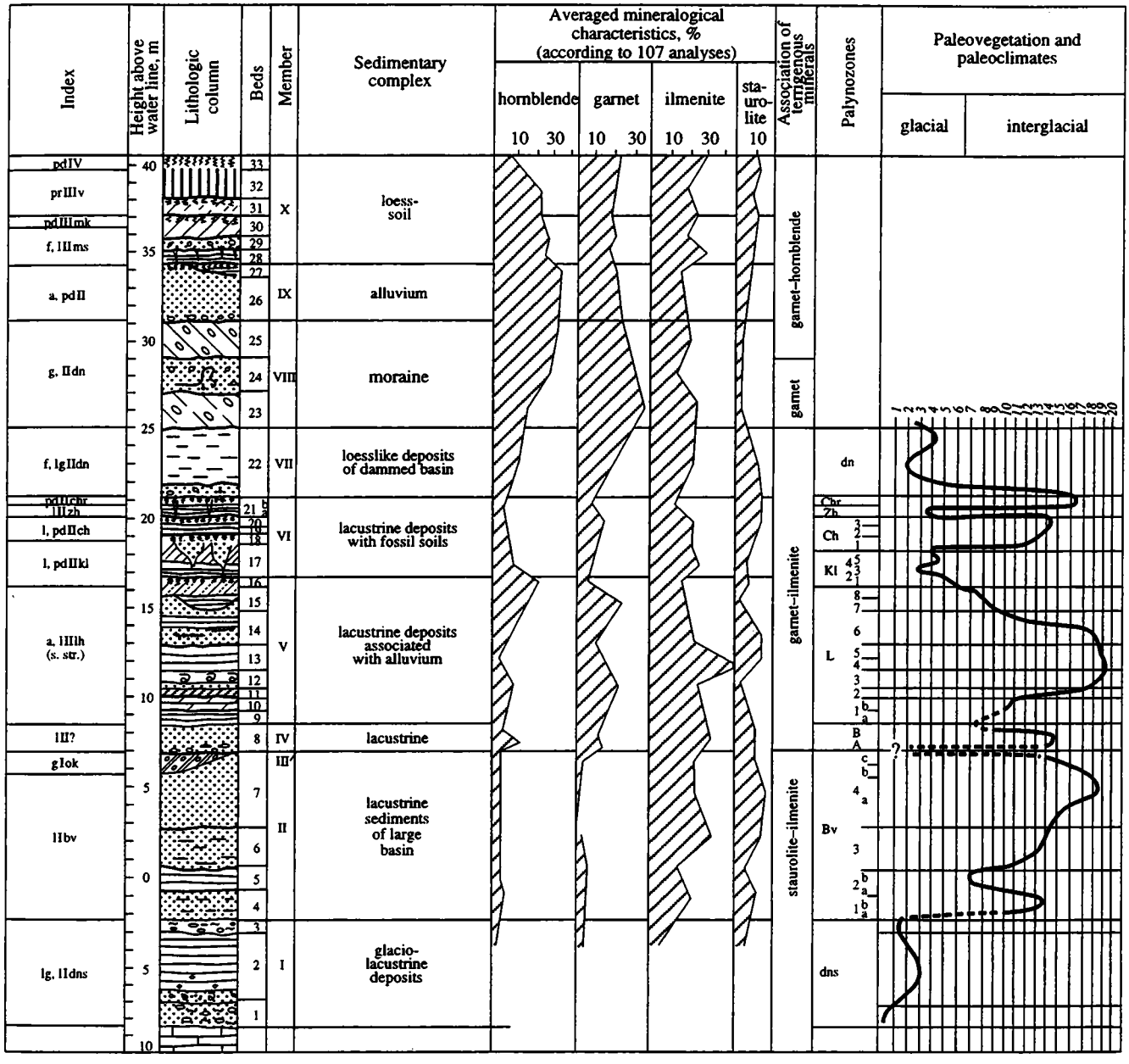

Fig. 3. The Pleistocene stratigraphic subdivisions and paleogeographic stages of the Likhvin stratotype region.

For column symbols see Fig. 1.

Paleovegetation and paleoclimates: (*1*) ice sheet; (2) periglacial tundra; (*3*) periglacial: forest-tundra; (*4*) periglacial forest-steppe; (*5*) light pine-birch forest; (*6*) light fir-pine-birch forest; (*7*) spruce forests; (*8*) birch forests with admixture of broad-leaved plants; (*9*) pine-birch forests with admixture of broad-leaved plants; (*10*) birch-pine forests with admixture of broad-leaved plants; (*11*) pine-spruce forests with admixture of broad-leaved plants; (*12*) birch and broad-leaved forests; (*13*) pine-cedar and broad-leaved forests; (*14*) pine-spruce and broad-leaved forests; (*15*) pine and broad-leaved forests; (*16*) pine-fir and broad-leaved forests; (*17*) broad-leaved (*Quercetum mixtum*) forests; (*18*) broad-leaved forests with dominating hornbeam; (*19*) spruce and broad-leaved forests with subtropical elements; (*20*) broad-leaved forests with subtropical elements.

Oka stratotype and Dainava moraine in Lithuania. Stratigraphic relationships between the Oka moraine and the fluvio lacustrine deposits of the reference section were revealed by a number of cuttings. The moraine accumulations and their subsequent erosion are indirectly evidenced by the composition of the deposits of the next stage of sedimentation.

The Likhvin Superhorizon (s. lato)

The pre-Likhvin deposits include the 2-m-thick distinctly lacustrine sandy Bed 8, the basal pebbles of which indicate a change in sedimentation environments and provenance of terrigenous minerals (Figs. 1–3). Big clasts are frequently composed of crystalline erratic rocks, and the garnet–ilmenite mineral assemblage has a significant admixture of exotic hornblende and other amphiboles and pyroxenes of Scandinavian origin (up to 14–16%). These features suggest the reworking of the moraine material, which was accumulated at the preceding stage and transported into the lake basin by river streams.

The taphonomic peculiarities of pollen and spore grains; abundance of fossil thermophilic plants; and occurrence of *Salvinia natans, Myriophyllum*, and other hydrophytes unambiguously indicate that Bed 8 (sands with gravel and pebble at the depth of 34.0–36.0 m) was accumulated in the interglacial period when coarse fragments of the alluvial material (products of glacial debris erosion) were transported into the shallow-water lake depression. It is possible that the change in the source areas of terrigenous minerals, which was established on a mineralogical and petrographic basis, as well as the influx of the exotic glacial material are not evident from the palynospectra of boundary deposits between Beds 7 and 8 because of the widely spaced sampling of well 152816. Like Bed 7, Bed 8 was accumulated at the time of spruce and broad-leaved forests when *Carpinus betulus* dominated among the thermophilic plants. The insignificant climatic deterioration is fixed by the spectrum from the upper part of Bed 7, which demonstrates the prevalence of lime (*Tilia cordata*) in the broad-leaved assemblages detectable among dominating mixed forest specimens. The comparative analysis of the floral composition shows that the disappearance of some early Pleistocene species and the first appearance of pollen of *Zelkova* sp., *Celtis* sp., and *Ulmus propinqua* in the section are confined to Bed 8. These taxa cannot specify the stratigraphic assignment of Bed 8, as they are the characteristic components of both the Likhvin and interglacial early Pleistocene palynofloras of the central Russian Plain. However, the disappearance of pollen of *Cedrus* sp., *Tilia amurensis* and spores of *Osmunda regalis, Woodsia manchuriensis*, and *W. fragilis* peculiar to the Belovezh'e palynoflora of the section may provide indirect evidence of the middle Pleistocene age of this bed. A period of its accumulation corresponds to palynozones A and B, which mark two phases in the vegetation development in the Upper Oka region:

A—zone of spruce, hornbeam, oak, elm, nettle tree, and dzelqua; phase of spruce–elm–oak–hornbeam forests with species of *Tilia, Celtis*, and *Zelkova*;

B—zone of spruce pine and broad-leaved plants; phase of spruce forests with an admixture of pine, European hornbeam, English oak, and elm (*Ulmus laevis, U. campestris*).

Based on the cooling trend recorded in the lower part of the overlying Bed 9, we related the sands with the gravel and pebble of Bed 8 to the first warming stage of the Likhvin interglacial (Bolikhovskaya, 1985). In view of the controversial lithological and palynological information concerning Bed 8, its stratigraphic position can be unambiguously defined after additional results.

The Likhvin Horizon (s. str.)

Of great interest is the lacustrine–alluvial complex (sedimentary complex V) representing the stratotype of the Likhvin Interglacial. This complex, up to 20 m in total thickness, is diverse in facial–genetic and lithologic aspects. A complicated combination of lacustrine, boggy, and alluvial deposits can be observed in the section and along the strike. Their spatial and age relationships are shown in Figs. 1 and 3.

The central part of the exposure exhibits the inherited lenses of basinal and oxbow deposits of different ages, which lie above a sharp erosion surface with angular disconformity. The axial parts of these lenses are displaced relative to one another. Individual beds vary in thickness over the area, and their total thickness amounts to 6–10 m. The base of the complex is 5 m above the water line, and stretches for about 200 m along the valley. Upsection, the following successive strata are noted: basinal rhythmically bedded sandy-clayey deposits of the periglacial type (Bed 9);[2] oxbow deposits: foliated marl containing abundant plant megafossils and fish imprints (Bed 10); bedded silts (Bed 11); dark-colored micaceous sands with molluscan shells (Bed 12); lacustrine-boggy green-blue clays with vivianite (Red 13); sandy sediments of the floodplain type, up to 6 m in thickness, including embryonic soils (Bed 14); clayey loam deposited near the valley in the oxbow, lake and bog settings (Bed 15), and overlain by a podzolic horizon of fossil soil FS_7. The distinguished facies variations are associated along the strike with the alluvial channel facies; the well sorted horizontally and cross-bedded light-colored sands, with interbeds of gravel and pebble.[3] In contrast to the Oka moraine, siliceous clasts dominate here over crystalline rocks (metamorphic and basic rocks, granite, quartz),

[2] They yielded *Dicrostonyx okaensis* (Aleksandrova, 1982).
[3] The pseudostratotype of the Oka moraine.

while low-resistant calcareous rocks are reduced to a minimum.

Palynoflora from the Likhvin stratotype consists of almost 90 taxa, 60 of which were identified at the species rank. The characteristic components, which are important for climatostratigraphic reconstructions and indicative of the flora type, vegetation and climate of the Likhvin Interglacial, include: *Tsuga canadensis, Taxus baccata, Larix* sp., *Abies alba, Picea* s. *Omorica, P. excelsa, Pinus* s. *Cembra, P.* s. *Strobus, P. sylvestris, Betula* s. *Costata, B. pendula, B. pubescens, Pterocarya fraxinifolia, Juglans cinerea, J. regia, Carpinus betulus, Castanea sativa, Fagus sylvatica, F. orientalis, Quercus petraea, Q.* cf. *castaneifolia, Q. robur, Q. pubescens, Zelkova* sp., *Celtis* sp., *Ulmus propinqua, U. laevis, U. campestris, Fraxinus* sp., *Tilia platyphyllos, T. tomentosa, T. cordata, Acer* sp., *Corylus colurna, C. avellana, Ilex aquifolium, Alnus glutinosa, A. incana, Ligustrina amurensis, Buxus* sp., *Rhododendron* sp., *Vitis* sp., *Myrica* sp., *Osmunda claytoniana, O. cinnamomea, Salvinia natans,* and others. In the younger interglacial palynofloras of the section, there is no pollen of hemlock, yew, wing-nut, oil nut, chestnut, holly, chestnut oak, or box, as well as spores of Clayton royalfern and other representatives of European, Mediterranean, East Asian and North American floras.

The upper part of Bed 9 related to the terminal phase of the cooling displays the palynozone of spruce and pine with single cryophytes (phase of pine–spruce forests with *Alnaster fruticosus, Betula fruticosa* and *B. nana* in the shrub layer reflects the endothermal cooling conditions).

The finer climatostratigraphic units of the Likhvin Horizon s. str. are represented by palynozones L1–L8, which are indicative of the major floral and vegetational successions at all stages of this long interglacial period (Figs. 2 and 3):

L1—zone of spruce, pine with single oak and lime: phase of birch–coniferous and coniferous forests:

1a—subphase of pine–spruce forests with admixture of oak, lime, and birch;

1b—subphase of birch–pine–spruce forests (with an admixture of elm) and alder groves;

L2—zone of spruce and pine with oak, elm, lime, hornbeam, and alder (pollen sum of forest-forming broad-leaved plants amounts to 4–5%): phase of pine–spruce forests with an admixture of broad-leaved plants:

2a—subphase of pine–spruce forests with an admixture of oak (prevalent) elm, lime, and hornbeam;

2b—subphase of pine–spruce forests with an admixture of hornbeam (prevalent), oak, elm, and lime;

2c—subphase of pine–spruce forests with an admixture of hornbeam, oak, elm, and lime in equal proportions, and the first appearance of *Juglans regia*;

L3—zone of spruce and broad-leaved plants (pollen sum of the latter is 12–29%): phase of spruce–oak and spruce–oak–hornbeam forests:

3a—subphase of pine–oak forests (with an admixture of pine, elm, and lime) and alder groves;

3b—subphase of spruce–oak–hornbeam forests with an admixture of fir, lime, elm, and others;

L4—zone of hornbeam, oak (their pollen sum is 40–72%) and alder: phase of oak–hornbeam forests and alder groves;

L5—zone of hornbeam, oak, fir, spruce, and alder with *Pterocarya fraxinifolia, Juglans cinerea, J. regia, Fagus sylvatica,* and *F. orientalis* (pollen sum of broad-leaved plants is 6–34%): phase of dark-coniferous and broad-leaved forests:

5a—subphase of spruce–fir, oak–hornbeam forests (with representatives of *Pterocarya fraxinifolia, Juglans cinerea, J. regia,* and other thermophilic and hydrophilic plants) and alder groves;

5b—subphase of fir–spruce and hornbeam–beech–oak forests and alder groves;

5c—subphase of cedar–spruce–fir and oak–lime–hornbeam forests and alder groves;

L6—zone of oak, beech, hornbeam, pine, and spruce with *Ilex aquifolium* (pollen sum of broad-leaved plants is 8–43%): phase of spruce–pine–hornbeam–beech–oak forests with representatives of *Ilex aquifolium*;

L7—zone of birch and pine with spruce and broad-leaved plants (pollen sum of the latter is up to 10%): phase of spruce–birch forests with an admixture of spruce, lime, and hornbeam;

L8—zone of birch with pine, spruce, and oak (singly): phase of birch forests with insignificant admixture of pine, spruce, and oak.

Prior to the comprehensive investigations of the Likhvin section initiated in 1968 by specialists from Laboratory of Recent Deposits and Pleistocene Paleogeography at the Moscow State University, the studies of floral and faunal remains by V.N. Sukachev, B.S. Dokturovskii, P.A. Nikitin, M.P. Grichuk, V.P. Grichuk, E.N. Ananova, P.I. Dorofeev, A.P. Jouse, V.D. Lebedev and others were devoted to palynological spectra from the oxbow sediments only (Beds 9–12 in Figs. 1–3). Therefore, the premature inference was made on the basis of the spore–pollen evidence that the Chekalin section exhibited the deposits of the first half of the Likhvian Interglacial, i.e., zones L1–L3 (Grichuk, 1989). A detailed palynological analysis (using the improved procedures for pollen and spore extraction) was applied to the entire pre-Dnieper sequence beneath the moraine. Most of the sequence was earlier considered to be paleontologically barren. It was established that the Likhvin Horizon (*s. str.*) is represented in the stratotype section in full scope (Beds 9–15 in Figs. 1–3; Bolikhovskaya, 1974, 1995). Moreover, the deposits of subsequent pre-Dnieper glacial and interglacial rhythms are present there as well.

The Likhvin–Dnieper Horizons

The Likhvin Interglacial deposits are overlain by Sedimentary complex VI, about 5 m in total thickness, which consists of lacustrine horizontally bedded predominantly clay and loam strata (Beds 17–21) occurring above the erosion surface. The sediments include the hydromorphic fossil soils FS_7, FS_6, FS_5, and FS_4 and the cryogenic horizons with permafrost textures and structures. Some hiatuses in sedimentation and repeated changes in the landscape and climatic environments are recorded. These stages in the lake development may be related to the wide floods alternating with the soil-forming processes in the dried terraces.

The palynological studies distinguished the deposits of interglacial and glacial genesis, which belong to four stratigraphic units. The names of two first horizons—the Kaluga Horizon corresponding to the cooling trend and the Chekalin Horizon corresponding to the warming trend—were suggested by Sudakova (1975), and the Zhizdra and Cherepet' horizons, corresponding to glacial and interglacial rhythms respectively, were named by Bolikhovskaya after the nearest toponyms—the Zhizdra River and the Cherepet' Settlement.

The Kaluga Horizon

In the Likhvin section, this horizon is represented by lacustrine and alluvial sandy sediments overlain by fossil soil FS_7 and by parent rock FS_6, which has a mottled structure with frost fissures and ground wedges (Beds 16, 17). The characteristic components of the palynoflora of the Kaluga cooling trend are *Larix* sp., *Alnaster fruticosus, Betula fruticosa, B. nana, Dryas octapetala, Selaginella sibirica, Lycopodium appressum, L. pungens* (*L. dubium*) and other cryophytes. The climatostratigraphic subdivisions of the horizon are expressed in palynozones k11–k15, whose number and high diversity suggest the long duration of the Kaluga cooling and the expansion of the ice sheet up to the northern areas adjacent to Chekalin.

k11—zone of birch, pine, and larch: phase of larch–pine–birch forests;

k12—zone of birch and pine with participation of cryophilic species: phase of light pine–birch forests with the shrub layer of *Juniperus, Betula fruticosa*, and *B. nana;*

k13—zone of birch and cryophytes: phase of periglacial birch forest-tundra of the cryohygrotic substage of the ice sheet, which likely occured in higher latitudes;

k14—zone of pine, birch, and cryophytes: phase of light pine–birch forests (interstadial?);

k15—zone of birch and cryophytes: phase of periglacial forest-steppes at the cryoxerotic substage of the ice sheet; areas of light birch forests and shrub formations consisting of *Alnaster fruticosus, Betula fruticosa*, and *B. nana*, and open areas with a grass-shrub cover composed of heathoid, sedge–grass–herb and club–moss–fern associations.

The Chekalin Horizon

This horizon comprises the podzolic horizon A_2 of the fossil soil FS_6, lacustrine clays, and the fossil soil FS_5 (brown parasoil) (Beds 18–20). The characteristic components of the Chekalin interglacial palynoflora include *Picea* s. *Omorica, P. excelsa, Pinus*. s. *Cembra, P. sibirica, Carpinus betulus, Quercus robur, Tilia cordata, T. platyphyllos, T. tomentosa, Acer* sp., *Ulmus laevis, U. glabra,* and *U. campestris*.

Because the upper part of FS_5 is eroded, the terminal phases of the Chekalin warming are not pronounced in the section. Palynozones Ch1–Ch2 distinguished in the spore–pollen diagram represent the following climatostratigraphic zones and phases in the paleophytocoenosis development.

Ch1—zone of birch and pine with spruce, elm, and oak: phase of pine–birch forests with an admixture of spruce, elm, and. oak;

Ch2—zone of lime, elm, oak, spruce and pine with hornbeam: phase of coniferous broad-leaved forests with an admixture of birch, alder, and hornbeam:

2a—subphase of pine–spruce and oak–elm–lime forests;

2b—subphase of pine–spruce and elm–lime–oak (with hornbeam) forests;

2c—subphase of pine–spruce and elm–oak forests;

Ch3—zone of oak, elm, lime, and spruce with maple and pine: phase of mixed spruce–lime–elm–oak forests with an admixture of pine, birch, alder, and maple.

The Zhizdra Horizon

In this section it is represented by lacustrine–boggy clays of the lower half of Bed 21 and by the wedges cutting the top of the underlying FS_5 to the depth of 0.70 m. According to the spore–pollen data, the horizon corresponds to a single climatostratigraphic zone (palynozone zh in Fig. 2):

Zh zone of treelike birches, *Betula fruticosa*, alder groves, and other cryophytes: phase of periglacial forest-tundras.

In contrast to similar landscapes of the previous cold Kaluga period, the periglacial forest-tundras of the Zhizdra cooling episode include a lesser amount of cryophilic plants but more xerophytes. The characteristic components of the Zhizdra palynoflora are *Pinus sylvestris, P. sibirica, Juniperus, Betula pubescens, B. pendula, B. fruticosa, B. nana, Alnaster fruticosus, Salix* sp., *Ephedra* sp., *Cannabis sativa, Artemisia* s.g. *Seriphidium, Botrychium boreale, B. simplex,* and *B. virginianum*.

The Cherepet' Horizon

This horizon corresponds to the boggy-gley soil FS_4. Finer climatostratigraphic subdivisions and successions of the interglacial forests which dominated during the Cherepet' warming are represented by palynozones Chr1–Chr5 (Figs. 2 and 3).

Chr1—zone of birch and pine with oak and hop-hornbeam: phase of pine–birch forests with an admixture of oak and hop-hornbeam in the undergrowth;

Chr2—zone of oak, hornbeam and birch: phase of hornbeam–oak forests with a lime admixture and birch forests;

Chr3—zone of pine and birch with elm and willow: phase of birch–pine forests with an admixture of elm and willow groves (the endothermal);

Chr4—zone of birch, Scotch pine, Mexican stone pine, oak, and elm: phase of pine–cedar–birch and elm–oak forests and willow groves;

Chr5—zone of birch and pine with single broad-leaved plants: phase of pine–birch forests with an admixture of elm and lime.

The characteristic components of the Cherepet' interglacial palynoflora include *Pinus* s. *Cembra, Carpinus betulus, C.* cf. *orientalis, Ostrya* sp., *Quercus robur, Q.* cf. *pubescens, Tilia cordata, T. tomentosa, Ulmus laevis*, and *U. campestris*.

The Dnieper Horizon

The Dnieper glacial sedimentation (sedimentary complexes VII, VIII) started at the time of intense flooding of this area, when a thick sequence (from 3 to 12 m) of loesslike fluvio-glacial bedded sandy loam and loam of greenish, pale and yellow color carbonates (Bed 22) were accumulated in the dammed semiflowing basins. The deposits are widespread laterally, and yielded the lemming fauna (Agadzhanyan, 1973) and tundra-steppe palynospectra. The latter suggest the predominance of the arctic-boreal shrub communities (*Betula fruticosa, B. nana,* and *Alnaster fruticosus*), and also include representatives of grass-sedge and green moss bogs along with specimens typical of eroded and locally turfed ecotopes occupied by sagebrush associations with *Artemisia* s.g. *Seriphidium, Ephedra distachya, Thalictrum Caryophyllaceae, Cichoriaceae, Selaginella sibirica, Lycopodium pungens*, and others.

The Dnieper moraine has the total thickness of 10–15 m and a three-member structure. The upper and lower units, which are represented by boulder-loamy sands and loams, are separated with boulder sand and differ in color, structural peculiarities, material composition, and magnetic properties (*Putevoditel'* ..., 1981). As compared with the Oka moraine, the Dnieper Horizon is characterized by an increased amount of igneous rocks among the clastic inclusions (20–30%), by the garnet-hornblende association of terrigenous minerals (20–30% each), and by the submeridional clast orientation. The upper age limit of the Dnieper moraine is evident from the radio-thermoluminescent data (Vlasov *et al.*, 1981) for the alluvium overlying the moraine (210 ± 50 ka, L-9) and the fossil soil FS_3 above the alluvium (250 ± 60 ka, L-10). These values and the position of the Dnieper moraine suggest that the Dnieper glaciation spaned the interval from 380 to 290 ka and proceeded in three stages.

The Dnieper moraine is overlain with erosion marks by the buried alluvial deposits (sedimentary complex IX) enclosing the fossil soil FS_3 (Sudakova, 1973). The section is completed with the blanket loesslike sediments containing the Mezin and Bryansk fossil soils.

CONCLUSION

1. The Don, Oka and Dnieper glacial horizons were distinguished in the Likhvin reference section. The Don Glacial corresponds to the morainelike and lacustrine–glacial deposits accumulated under conditions of the severe climate of the periglacial forest-tundra. The glacial Oka Horizon was recognized, and its stratigraphic relationships with the Belovezh'e and Likhvin horizons were defined. The marker horizon of the three-member Dnieper moraine rests on the complex Oka–Dnieper sequence, corresponding to the so-called "Great Likhvin" event.

2. The lacustrine sediments of the interglacial Belovezh'e Horizon of the Likhvin section display specific palynological characteristics, such as the predominance of mixed spruce–broad-leaved forests in the optimal phytocoenosis phases and the combination of indicative taxa such as *Cedrus* sp., *Rhus* sp., *Tilia amurensis, Osmunda regalis, Woodsia manchuriensis,* etc.

3. Contrary to existing opinions, the interglacial Likhvin Horizon (s. str.) is completely represented in the Likhvin stratotype section. It is of a complex climatostratigraphic structure: eight stratigraphic units (L1–L8); two climatic optima and the endothermal, as in the Belovezh'e Interglacial Horizon; and a specific assemblage of various thermophilic exotics (*Taxus* sp., *Pterocarya fraxinifolia, Juglans cinerea, Castanea sativa, Ligustrina amurensis, Buxus* sp., and others).

4. It is suggested that the deposits of the Oka–Dnieper interval, which comprises three interglacial horizons and two glacial horizons representing together the "Great Likhvin" event, should be united into the Likhvin Superhorizon of the regional stratigraphic scheme for the central Russian Plain, because the Kaluga Cryochron, Chekalin Themochron, Zhizdra Cryochron, and Cherepet' Thermochron after the Likhvin Interglacial (s. str.) are most likely of independent paleogeographic importance.

Thus, the significance of the Likhvin reference section for the Pleistocene stratigraphy, paleogeography, and correlation is evident from the following: it includes the representative Likhvin stratotype, the

underlying Oka moraine and palynologically characterized lacustrine Belovezh'e deposits, and displays the distinct stratigraphic position of the Dnieper moraine, and co-occurring glacial (up to three), cryogenic (up to five) and fossil soil (not less than seven) marker horizons, which can serve as a paleogeographic basis for interregional correlations between the Pleistocene sequence in the central Russian Plain and those in the northern glacial provinces and southern loess areas.

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BRIEF COMMUNICATIONS

The Terminal Ordovician Event and Paleozoogeography of Benthic and Planktonic Faunas (Brachiopods and Graptolites)

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Abstract—Climatic changes that resulted from the Late Ordovician glaciation coincided with the mass extinction and drastic alterations of the paleozoogeography of marine biota. The extinction of warm-water faunas in epicontinental seas, as well as some brachiopod assemblages on the open shelf at the beginning of the Hirnantian, apparently resulted from the glacio-eustatic regression and decrease of water temperature in low latitudes favored by the concentration of major continents in the southern hemisphere. The Hirnantian was a time when new paleozoogeographic relations were formed, which extended the distribution areas of short-lived cold-water faunas of the temperate belt, namely, of the Hirnantian brachiopod assemblage. After their extinction at the beginning of the Rhuddanian, the vacant ecologic space was occupied by impoverished assemblages of warm-water brachiopods, namely, virgianids and atrypids, which survived the biocoenotic crisis. With the advance of Llandoverian transgression, they occupied epicontinental seas and assumed a cosmopolitan character. The provincial differences of Graptoloidea were substantially obliterated before the initial Ashgillian. After the extinction event in the Hirnantian and reduction of black-shale sedimentation areas, the impoverished post-crisis diplograptid assemblages developed. The beginning of the adaptive radiation of graptolites corresponded to the end of the Hirnantian; however, its global effect was manifested later in the Rhuddanian, when Graptoloidea once again assumed planetary distribution areas.

Key words: *mass extinction, biocoenotic crisis, regression, transgression, glaciation, Hirnantian, Rhuddanian, graptolites, brachiopods, paleogeography.*

The Late Ordovician glaciation and related climatic changes coincided with one of the most substantial rearrangements of marine biota. It was characterized not only by the mass extinction of biota, but also by the drastic change of paleozoogeography of both benthic and pelagic faunas. If, over most of the Ordovician a contrasting paleozoogeographic differentiation occurred, marine fauna in the Llandoverian became cosmopolitan. These changes took place mainly in the Hirnantian; the duration of the latter, according to different estimates, was 2 or 3 m.y. Such a time span was insufficient for any considerable change in the relative positions of continents, that could have resulted in the rearrangement of the system of oceanic currents formed in the Ordovician. It is more plausible that climatic changes caused by the Late Ordovician glaciation triggered a number of realignements in the paleozoogeography of benthic and, to a lesser degree, pelagic faunas.

Before proceeding directly to an overview of the paleogeography of the Late Ordovician and Early Silurian fauna, we should briefly mention the distribution of large continental and water masses at the Ordovician–Silurian boundary. At that time, according to paleotectonic reconstructions (Scotese and McKerrow, 1991), major continents were located in the southern hemisphere. The largest landmass, Gondwana, extended from the equator to the South Pole, with its African and South American parts located mainly within the southern polar circle. It was separated by the sublatitudinal Paleotethys ocean from the European (Baltia) and Siberian continents. The Iapetus Ocean separated the Siberian Plate and Baltia from North America (Laurentia). Most of the northern hemisphere was occupied by a vast oceanic basin, the Panthalassa Ocean.

The paleozoogeography of the Late Ordovician and Early Silurian faunas was most thoroughly studied by the example of brachiopods. The Ordovician belt zonality in their distribution was recorded (Rozman, 1977); therefore, the differentiation of brachiopod faunas at that time at least partially resulted from climatic differences. The high latitudes belt of cold-water faunas in the seas surrounding the Gondwana coasts was clearly distinguished for the Ashgillian. These were primarily brachiopod faunas of North Africa, Spain, and southern and central Europe, concentrated in the Mediterranean Province (Havlicek, 1989). The biota was of a low taxonomic diversity and simplified biofacies zonality. Orthids and, to a lesser degree, strophomenids were most characteristic of this province. At the same time, typical Late Ordovician brachiopod taxa, such as pentamerids, ancient atrypids, and trimerilids, were absent there. The paleozoogeography of the Ashgillian benthos fauna in temperate and low latitudes was more complicated due to the distribution of separated conti-

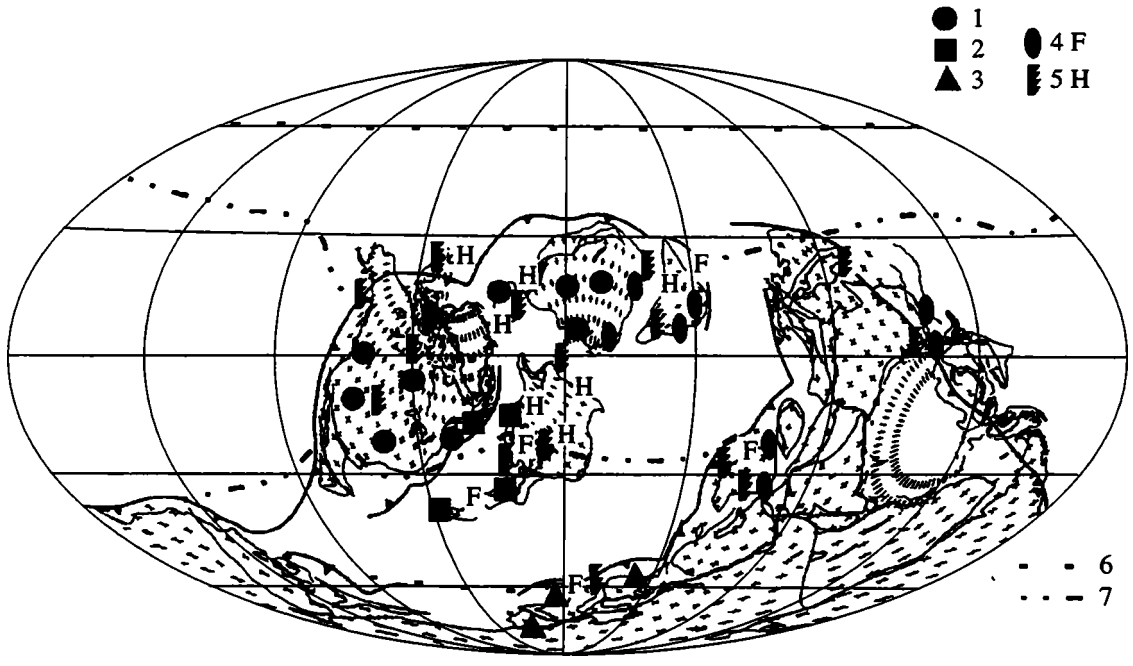

Fig. 1. Distribution of brachiopod biofacies and graptolites in the Early Ashgillian (the *supernus* Zone). Location of continents and lines of water mass convergence are according to Scotese and McKerrow, 1991, and Wilde, 1991.

(1–4) Brachiopod biofacies: (1) rhynchonellid–strophomenid (warm-water); (2) clitambonitid–orthid–strophomenid (temperate and subtropical latitudes); (3) enteletid (cold-water); (4) atrypid-plectombonitid and atrypid associated with corals *Agetolites* (warm-water); (5) localities of graptolite assemblages of the *supernus* Zone; F = the *Foliomena* assemblage (cold-water, the 5th benthic community) and H = the *Holorhynchus* assemblage (warm-water, the 3rd benthic community) (Rozman, 1977; Blodgett *et al.*, 1987; Havlicek, 1989); Convergences: (6) Polar; (7) Subtropical.

nental masses in these areas (Fig. 1). The distinguished main provinces mostly coincided there with the epicontinental and marginal seas, which bordered large landmasses (Rozman, 1977; Williams, 1973). Rhynchonellid-orthid-strophomenid assemblages were characteristic of Laurentia and Siberia, i.e., of the North American province or region (Williams, 1973). The peculiarity of the Baltic fauna (European province or region) was emphasized by the occurrence of clitambonitoids, whose diversity and abundance was distinctly lower than in the Middle Ordovician. The brachiopod faunas from the shelves of minor continents and island arcs, which clustered around the subequatorial Australian part of Gondwana, namely, southern China, Tarim, and Kazakhstan, had a somewhat different composition. In the Late Ordovician they were more taxonomically diverse and had the greatest biofacies differentiation. Ancient atrypids became numerous in these faunas for the first time, whereas trimerilids reached the highest diversity. The appearance of ancient virgianid assemblages with *Holorhynchus* and *Tcherskidium* was apparently associated with the shelves of island arcs and microcontinents along the periphery of the Panthalassa Ocean. During the second half of the Ashgillian, this fauna also penetrated into the Baltic paleobasin (Fig. 1).

Highly diverse assemblages of small lingulates were also typical components of the warm-water fauna from the temperate and low latitudes. They were dominated by acrotretids, which are known from the Ashgillian of Kazakhstan, Baltoscandia, and Ireland, and had a high taxonomic diversity up to the end of the Rawtheyan (Wright and McClean, 1991). The Hirnantian was the time of the termination of the Ordovician glaciation and formation of new paleozoogeographic relations. The latitudinal zonality of benthic faunas still occurred, and in some aspects, was more pronounced than in the Early Ashgillian, suggesting an increased temperature gradient from poles to the equator. The extended temperate belt was inhabited by the so-called Hirnantian fauna (Kosov province after Owen *et al.*, 1991). Following the cold-water currents along the western margins of Gondwana and Baltia, this fauna penetrated into subtropical and subequatorial latitudes (Fig. 2). In the extreme south, the seas surrounding the ice sheet located on the African and Arabian parts of Gondwana were inhabited by a sharply impoverished Hirnantian fauna with several endemic taxa. This provided the grounds for distinguishing the distribution area of this fauna as a separate Bani province (Owen *et al.*, 1991). The abruptly decreased distribution areas of warm-water faunas of the Edgewood province included minor epicontinental basins and marginal seas of Laurentia, Baltia, and northeastern Russia (Fig. 2).

Apparently, the relict warm-water Ordovician faunas could also occur on insular shelves of the marginal

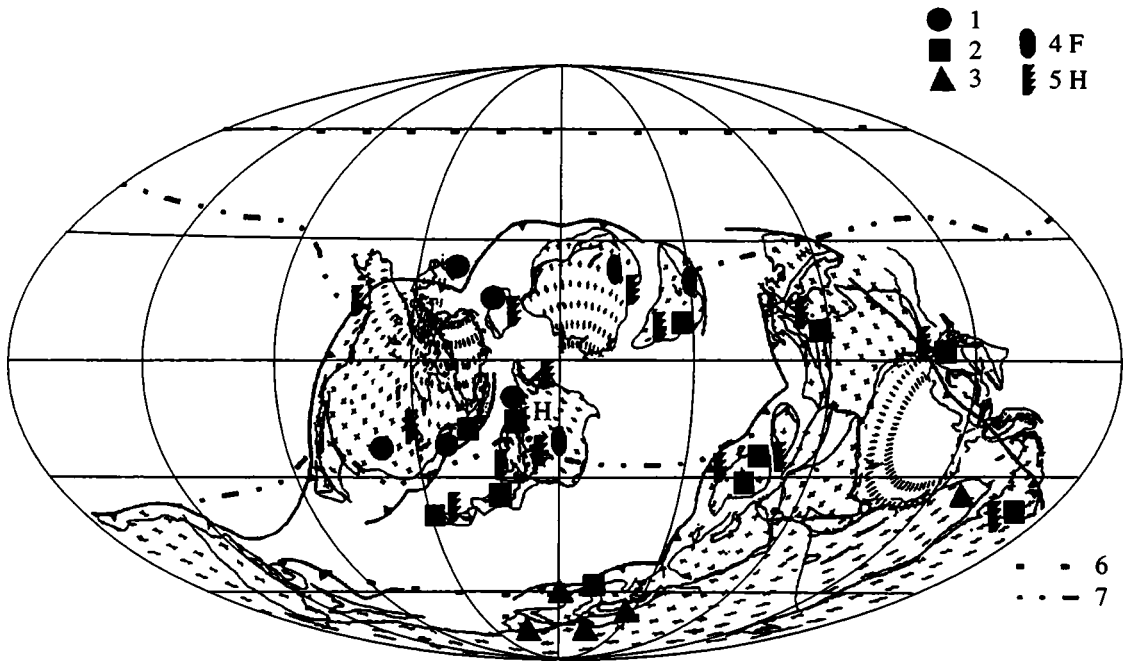

Fig. 2. Distribution of brachiopod biofacies and graptolites in the Late Ashgillian (Hirnantian).
(1–3) Brachiopod provinces: (1) Edgewood (warm-water); (2) Kosov (Hirnantian biofacies); (3) Bani (impoverished Hirnantian biofacies); (4) endemic brachiopod assemblages including relicts of ancient local fauna; (5) localities of graptolite assemblages of the *extraordinarius* and *persculptus* zones. H = *Holorhynchus* brachiopod assemblage. Convergences: (6) polar; (7) subtropical.

seas along the Panthalassa periphery. This is evidently the only explanation for the sudden distribution of some "Lazarus" taxa of the Rhuddanian reported from the Early Ashgillian, but unknown in the Hirnantian.

Comparison of the paleozoogeography of benthic fauna in the Early and Late Ashgillian (Hirnantian) showed that major changes were manifested in the substantial widening of distribution areas of relatively cold-water fauna of the temperate belt, which penetrated in some areas into the subequatorial regions, while, in contrast, areas of warm-water tropical fauna considerably decreased. The mass extinction of taxa associated with biofacies of the open shelf, such as those of *Foliomena* and microscopic lingulates, as well as of faunas from epicontinental seas, resulted in a considerable simplification of the biofacies zonality.

At the end of the Ashgillian, the glaciation terminated and the Hirnantian fauna was almost completely extinct. The vacant ecologic space began to be inhabited by a small amount of taxa that passed from the Ordovician and were components of warm-water faunas of Laurentia and Baltia. Following the Early Llandoverian transgression, this fauna entered epicontinental seas and became cosmopolitan in the Rhuddanian (Owen *et al.*, 1991; Stanton *et al.*, 1990). Virgianids and atrypids were its most characteristic members. The new biofacies zonality only began formation in the Rhuddanian, when it was not still quite pronounced. In the second half of the Rhuddanian, the radiation effect was already perceptible, and neoendemics appeared, such as the first spiriferoids known from Australia, southern China, and Kazakhstan.

The provincial differences in Early Ordovician graptolite faunas were already insignificant in the Caradocian. From the beginning of the Ashgillian, the taxonomic differences of still considerably diverse graptolite faunas became poorly pronounced and did not reach the provincial rank. It is suggested that in the Ashgillian, the areas inhabited by graptolites occurred mainly in tropical latitudes. However, all of the peculiar Ashgillian pelagic communities were almost completely eliminated before the beginning of the Hirnantian as a result of the mass extinction (Koren', 1991; Koren' and Rickards, 1979; Melchin and Mitchell, 1991). This event coincided with the reduction of accumulation areas of black-shale sequences and the decrease of graptolite distribution areas (Figs. 1, 2).

In the graptolite evolution, this period corresponded to the development of impoverished assemblages typical of postcrisis intervals. Radiation, starting from a single species of *Normalograptus,* locally began at the end of the Hirnantian (the *persculptus* Zone), but its global effect was manifested later, with the beginning of the vast Rhuddanian transgression (the *acuminatus* Zone). The Early Rhuddanian graptolite fauna occupied planetary distribution areas, and its taxonomic and morphologic diversity rapidly increased without pronounced signs of paleozoogeographic differentiation.

The Late Ordovician glaciation was associated with one of the greatest Phanerozoic crises. The strong influ-

ence of climatic changes on the marine biota apparently resulted from the concentration of major landmasses in the southern hemisphere, where the glaciation occurred (Scotese and McKerrow, 1991).

The disposition of continents, especially of the Australian part of Gondwana (Wilde, 1991), favored the penetration of cold surface waters into the subequatorial regions and, as a result, the wide distribution of relatively cold-water Hirnantian fauna. The spatial arrangement of continents and oceans at the Ordovician–Silurian boundary also contributed to the emergence of cold oxygen-saturated bottom waters, which formed around the southern polar ice sheet in low latitudes. The aeration of oceanic waters at middle depths strengthened. These processes resulted in the destruction of the system of oceanic water stratification formed in the Ordovician, and in the degradation of ecosystems associated with oxygen-depleted layers widely distributed in the southern hemisphere.

Graptolites were important components of Late Ordovician pelagic ecosystems. The vertical water circulation, which was intensified during the Ashgillian glaciation, apparently changed the traditional graptolite biotopes considerably and, thus, banefully influenced the diversity of assemblages. There is a correlation between the dynamics of the graptolite diversity and the decrease or expansion of black-shale sedimentation areas (Koren', 1991; Melchin and Mitchell, 1991). These facies serve as indicators of climate and oceanic bioproductivity throughout the whole Phanerozoic. Graptolites most likely inhabited layers poor in oxygen and rich in nitrates, which occurred between the oxygen-saturated surface waters and deeper area of the globally distributed oxygen-depleted zone. The Late Ordovician cooling resulted in the reduction of the area of oxygen-deficient waters, which rose up and were mixed. All of this activity, together with the reduced tropical belt and increased convection, could have led to the diversity decrease and complete elimination of the graptolite provincialism in the Late Ordovician. During the maximum phase of glaciation at the Hirnantian boundary, the stratification of anaerobic and disaerobic layers was completely eliminated due to the vertical and horizontal displacement of cold oxygen-rich waters to surface layers in tropical basins (Melchin and Mitchell, 1991).

The Hirnantian black-shale sedimentation in the south Chinese plate, located at that time close to the Australian part of Gondwana, can be explained by the presence of local barriers and seasonal upwelling near its western border (Wilde, 1991). As a result, up to the terminal Ashgillian, including the Hirnantian, this was a region of highly productive zones favorable for the development of pelagic graptolite assemblages that retained taxonomic diversity unusual for the Hirnantian. It was precisely this area where the radiation and formation of almost all the morphotypes of the new Silurian graptolite fauna began, even in the Hirnantian, prior to the global transgression in the Early Rhuddanian, when the oceanic water stratification was regained.

The eustatic regression at the Hirnantian boundary was baneful for the diverse and highly differentiated biota of epicontinental seas, especially for warm-water faunas of carbonate platforms. The decrease of water temperatures over vast territories could have disrupted the vital cycles of a great deal of pelagic and benthic organisms. There are reasons to believe that this factor substantially influenced the plankton composition and resulted in the extinction of many conodont and lingulate taxa at the Rawtheyan–Hirnantian boundary. We should also note that the simple trophic structure of Ordovician benthic communities dominated by filter feeders made them more vulnerable to unfavourable environmental changes than the Mesozoic and Cenozoic benthos (Sepkoski and Sheehan, 1983).

Thus, the rearrangement of the paleozoogeography of brachiopod faunas at the Ordovician–Silurian boundary resulted from the biotic crisis caused by climatic changes during the maximum development of the Late Ordovician glaciation. The extinction of warm-water faunas in epicontinental seas, together with some brachiopod assemblages characteristic of the open shelf at the beginning of the Hirnantian (Blodgett *et al.*, 1987), was apparently caused by the glacio-eustatic regression and oceanic water temperature decline in low latitudes, which was favored by the concentration of major continents in the southern hemisphere.

Causes of the extinction of the Hirnantian fauna in the expanded temperate belt during the glacial maximum are not clear. The cosmopolitan character of the Rhuddanian brachiopod fauna apparently resulted from the distribution of impoverished warm-water brachiopod assemblages, which survived the biocoenotic crisis, in the vacant ecologic space.

In the Ashgillian, graptolites played an important role in pelagic communities of low latitudes. The provincialism of the graptolite fauna, pronounced in the first half of the Ordovician, was almost eliminated by that time. The Hirnantian evolutionary crisis of the group can be explained by the reduction of the tropical belt and disruption of the oceanic water circulation formed in the Ordovician. The influence of climatic changes at the end of the Ordovician on the pelagic marine biota as a whole requires further investigation, because such important components as radiolarians and acritarchs are still insufficienty studied for a comprehensive analysis.

Reviewer N.M. Chumakov

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BRIEF COMMUNICATIONS

Cephalopod Limestone Facies of the Silurian: the Taimyr Peninsula and Carnic Alps

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Key words: *cephalopod limestones, Silurian, cephalopod and bivalve mollusks, Taimyr, Carnic Alps.*

Cephalopod limestones are spread worldwide in the Silurian deposits. They are reported from the Llandoverian of eastern Siberia (Myagkova, 1967; Bogolepova, 1989, 1990; Bogolepova and Kříž, in press) and China (Chen *et al.*, 1981), and from the Wenlockian and Ludlovian in the Czech Republic, Yugoslavia, Spain, Germany (Kříž, 1979, 1992), France (Babin, 1966), Austria (Schönlaub, 1980, 1985, 1992), Morocco (Termier and Termier, 1950), Algeria, Turkey (Pojeta *et al.*, 1976), Sardinia (Serpagli and Gnoli, 1977; Gnoli *et al.*, 1979; Ferreti, 1989; Gnoli and Serpagli, 1991), the Caucasus (Yanishevskii, 1918), southern Urals (Ivanov and Puchkov, 1984), Tien Shan (Volkova and Chernova, 1961), Novaya Zemlya (Nekhorosheva, 1981), Arctic Canada (Pojeta and Norford, 1987), and Taimyr. These rocks are of great importance with respect to both stratigraphy, because they are confined to narrow stratigraphic intervals, and paleogeography, because their distribution at those periods was controlled by near-equatorial surface currents. Most of the Silurian cephalopods probably preferred tropical environments. After the death of mollusks, their empty shells, floating like buoys, were transported by the surficial currents and accumulated at sea bottom. This latter process was influenced by temperature changes, current velocity, and the existence of

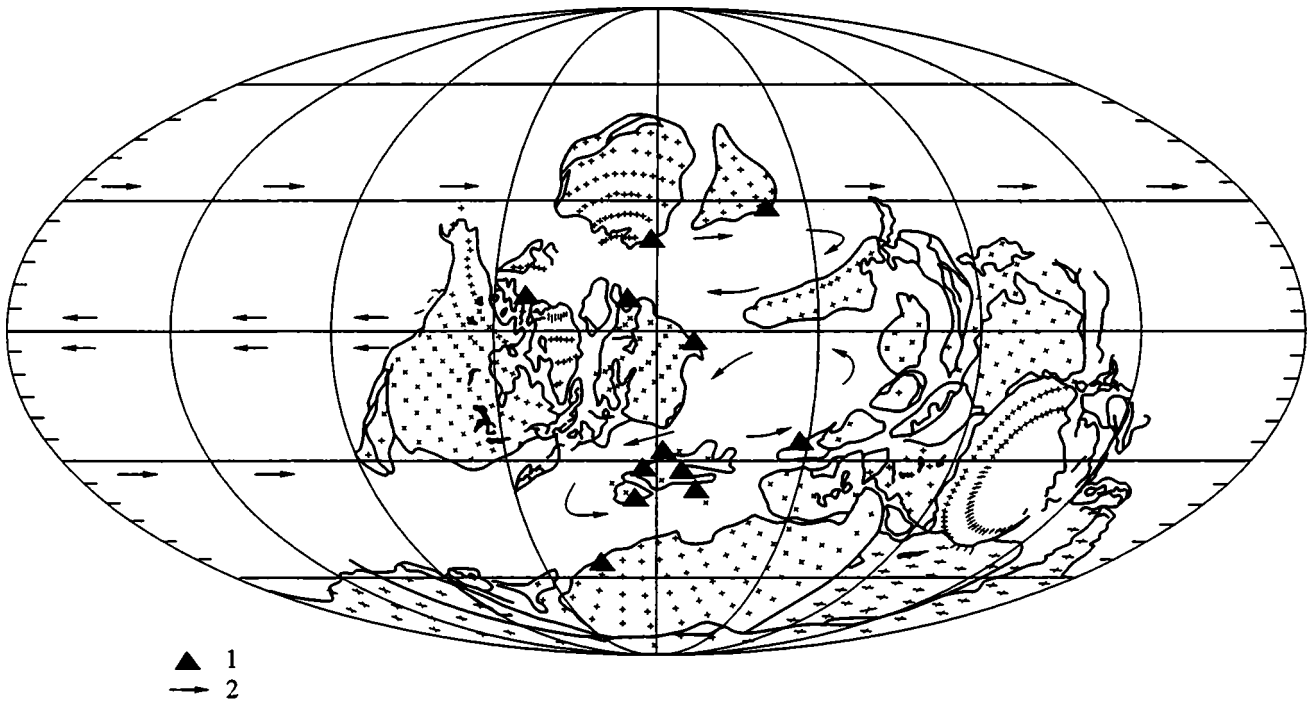


Fig. 1. Distribution of cephalopod limestone facies in the Late Silurian: (1) cephalopod limestones; (2) circulation system. Paleogeography is after (Scotese and McKerrow, 1990; Schönlaub, 1992).

shoals on their way. The cephalopod facies are represented by different color limestones: rose (the Southern Urals and Tien Shan), gray (the Caucasus and Czech Republic), and black (Sardinia, Novaya Zemlya, and Taimyr). They have variable thicknesses from a few centimeters in particular lenses to meters in thicker beds. The rock consists entirely of cephalopod shell remains, which usually have a certain orientation. The orientation patterns of orthoconical cephalopods are very important for reconstructing the directions of marine currents and their nature, as well as the paleogeographic setting of the basins and their links. For instance, the analysis of data (rose-diagrams), which I obtained while measuring the orientation of orthoconical nautiloid shells in the Rauchkofel section (the Carnic Alps), showed that local paleocurrents were directed from southwest to northwest in the Silurian, and from north-northwest in the Early Devonian. This conclusion is supported by paleotectonic, lithologic, and paleontologic data. In addition to numerous nautiloid of the Silurian orders, the biota of cephalopod facies included bivalve mollusks, ostracodes, rare trilobites, gastropods, and brachiopods.

Accumulation environments of cephalopod limestones were most favorable for the bivalve mollusks Cardiolidae. This family is known only from the Silurian. Its representatives bear distinct morphological features, which readily distinguish them from other Silurian or older bivalve mollusks. The cardiolid evolution is closely related to the evolution of orthoconical cephalopods and to the deposition of cephalopod limestone facies. The oldest cardiolids are known from the lower Wenlockian (the *centrifugus* Zone) of Bohemia, the Carnic Alps, and Wales. The maximum diversity of this family is characteristic of the Ludlovian. The stratigraphic interval of some cardoilid species is as narrow as one or several graptolite zones (Kříž, 1979), and this group is important for the subdivision and correlation of Silurian deposits.

The pelagic ostracodes Entomozoacea (Myodocopidae), whose stratigraphic and correlative potential as also high, are caracteristic of cephalopod biofacies. They are recorded in the Silurian deposits of England, Scotland, Sardinia, Bohemia, Poland, the United States, Canada, Australia, China, the eastern and western slopes of the Urals (Rozhdestvenskaya, 1979), and Taimyr. Occurrences of entomozoid and other pelagic faunas in black cephalopod limestones (Mediterranean, Taimyr) and their association with benthic organisms in light reefal, biohermal, and biostromal limestones (England, the Baltic–Scandinavian region, Bohemia) provide a great opportunity to correlate these different facies.

In the Middendorf Cave section located in the northwestern Taimyr Peninsula, cephalopod biofacies are confined to thin (about 20 cm) layers of black limestones incorporated into a thick member of graptolitic shales of the Middendorf Formation (Fig. 2). Nautiloid shells are small to medium in size, and distinctly ori-

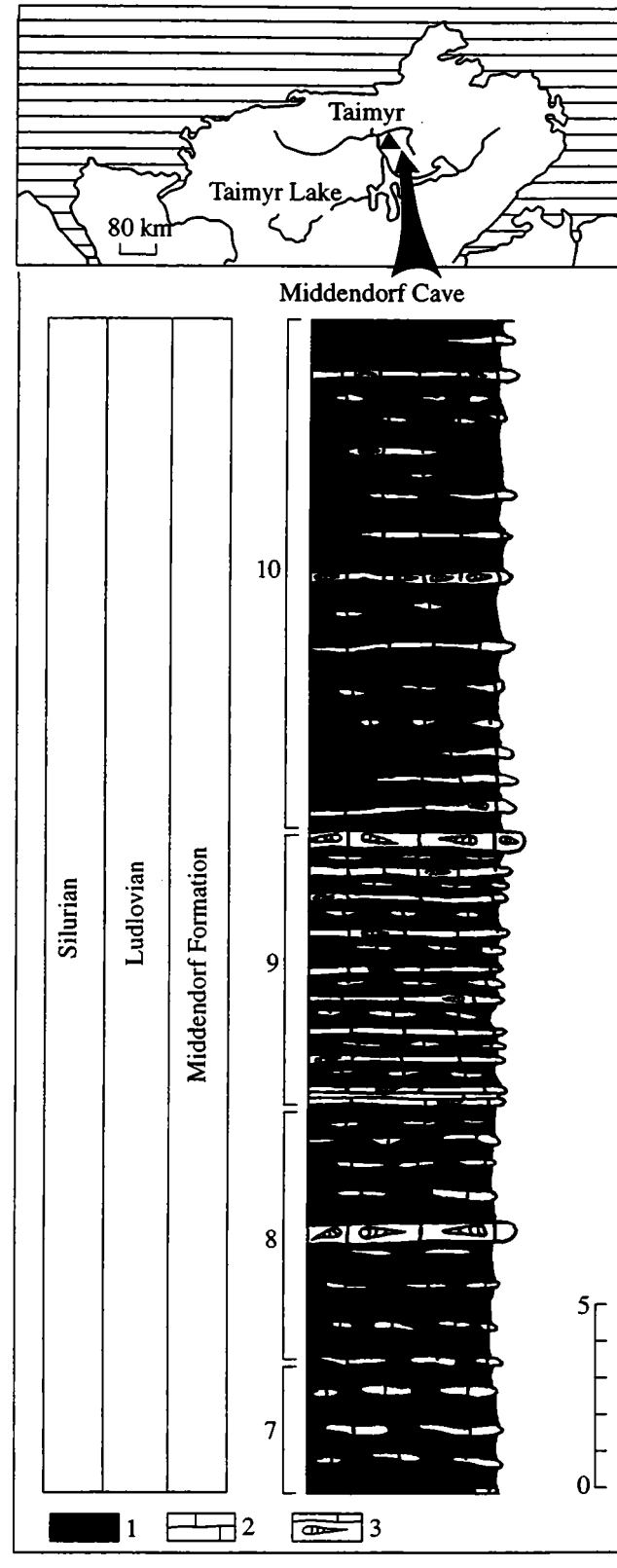
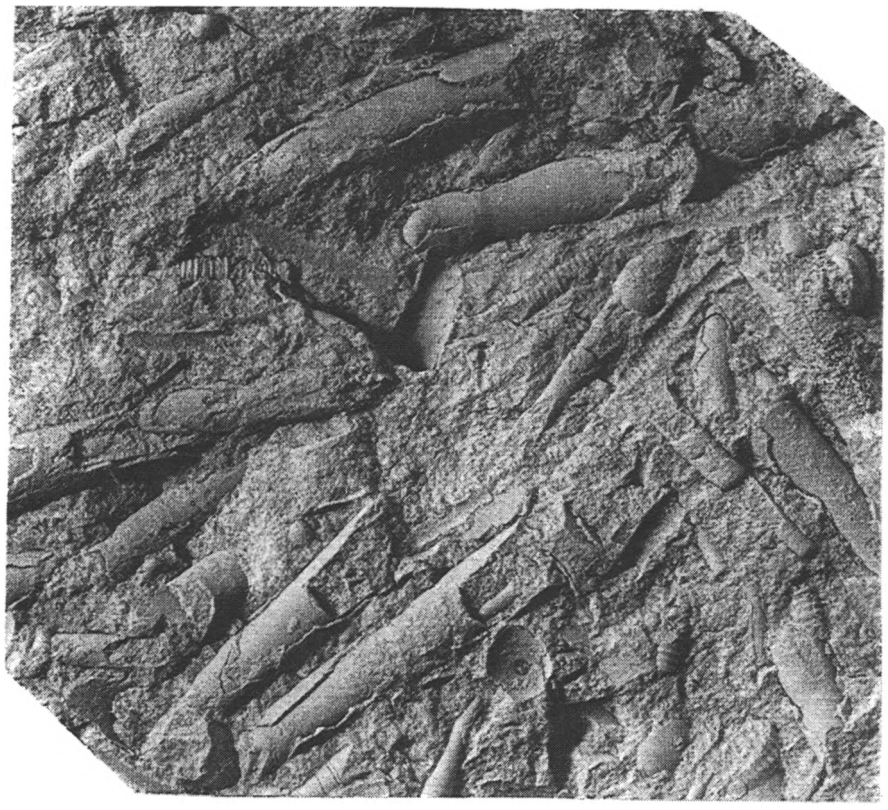
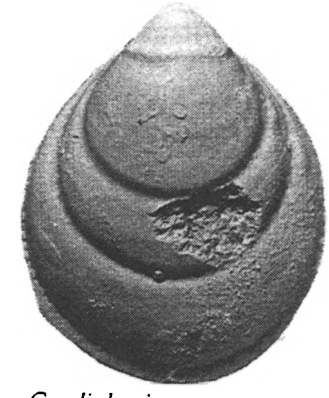


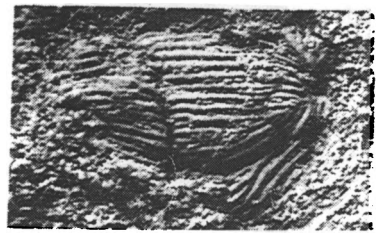
Fig. 2. Location of the section P-90218 (Middendorf Cave, Taimyr) and its fragment showing the stratigraphic position of cephalopod limestones (field observations by N.N. Predtechenskii): (1) black shales; (2) limestones; (3) cephalopod limestones.



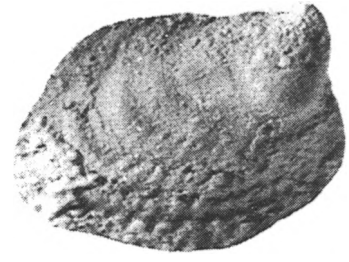
Cardiola aff. *consangius*



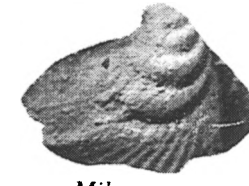
Cardiola signata



"Entomis" (R) migrans



Butovicella medea



Mila gyrans

Fig. 3. Fossils typical of the Upper Silurian cephalopod limestone facies of Taimyr: cephalopod limestone composed of *Michelinoceras* sp., *Hemicosmorthoceras semiannulatum* (Barrande), and *Anaspyroceras pseudocalamiteum* (Barrande), ×5; *Cardiola* aff. *consangius* Barrande, ×6; *Cardiola signata* Barrande, ×6; *Butovicella medea* (Barrande), ×13.5; *Mila gyrans* (Barrande), ×15; *"Entomis"* (r.) *migrans* Barrande, ×13.

ented. Faunal remains (Fig. 3) are represented by the abundant cephalopods *Hemicosmorthoceras semiannulatum, Geisonoceras* sp., *Kionoceras* aff. *doricum, Akrosphaerorthoceras* aff. *gregale, Michelinoceras* sp., *Anaspyroceras pseudocalamiteum* and *Parakionoceras originale*; bivalves *Cardiola signata, Cardiola* aff. *consangius, Dualina*? faba, *Mila gyrans, Maminka* cf. *comata*, and *Butovicella medea*, conodonts *Ozarkodina inclinata inclinata* and *Ozarkodina inclinata posthamata* (bed 8); ostracodes *"Entomis" (R) migrans* (bed 10), rare trilobites, and gastropods. Analysis of the stratigraphic distribution of the above-listed species allows the correlation of the host deposits (beds 8–10) with the *ploeckensis* Zone.

In the Carnic Alps, cephalopod limestones were studied in the Rauchkofel Boden section (Fig. 4). In this locality, they are thicker (about 2 m) gray and dark-gray beds of the Kok and Cardiola formations. Nautiloid shells are variable in size, from small to large, and uniformly oriented, which suggests the current stability during the long (Wenlockian to Pridolian) period (Schönlaub and Bogolepova, 1994). I determined from the upper Kok Formation (bed 2) the following cephalopods: *Merocycloceras declevis, Kionoceras* aff. *doricum, Hemicosmorthoceras laterculum,*

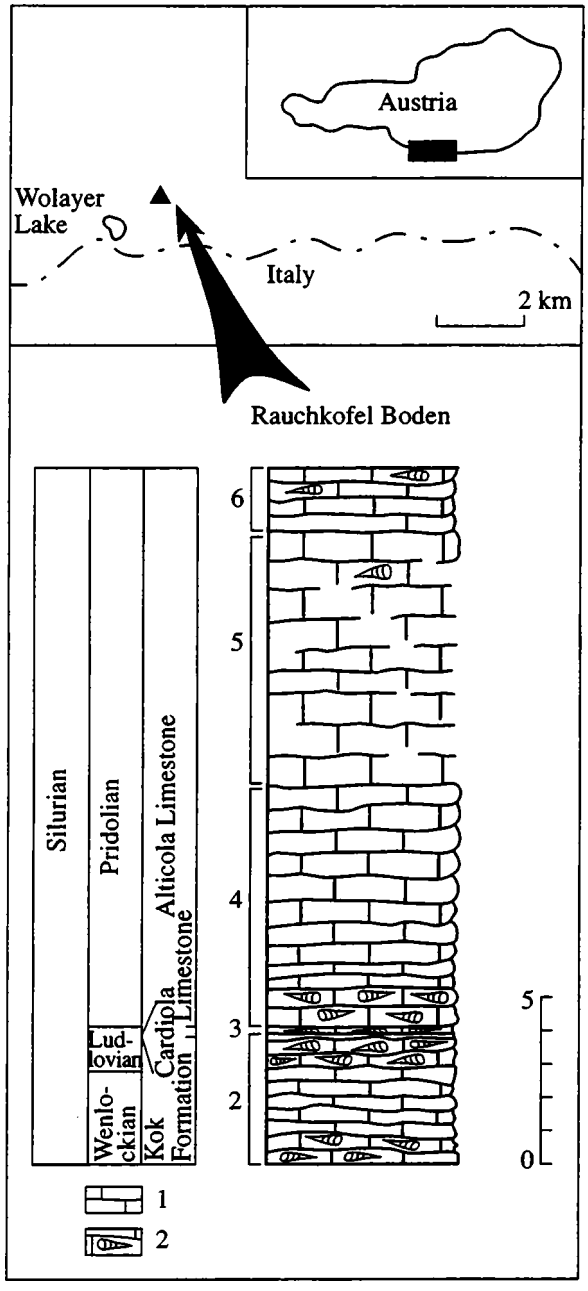


Fig. 4. Location of the Rauchkofel Boden section (the Carnic Alps, Austria) and its fragment showing the stratigraphic position of cephalopod limestones: (1) limestones; (2) cephalopod limestones.

Parasphaerorthoceras sp., *Parakionoceras* sp., and *Sphaerorthoceras* sp. associated with the bivalve mollusks *Cardiola signata, Cardiola consangius,* and *Mila complexa* (Kříž, 1979), some trilobites and rare gastropods. In taxonomic composition, this faunal assemblage is close to that from the Middedorf Formation of Taimyr. Conodonts from the Cardiola Formation (bed 3) correlate it with the *siluricus* Zone (Schönlaub, 1985), and this suggests that cephalopod limestones of the underlying upper Kok Formation are confidently correlative with the *ploeckensis* Zone.

Thus, the regions, which were widely spaced in the Silurian, experienced the single event that resulted in cephalopod biofacies formation. Occurrence of cephalopod beds in the Taimyr sedimentation basin at the same stratigraphic level (the *A. ploeckensis* Zone) as in the Southern Alps suggests that both these basins were characterized by similar tectonic and eustatic environments, which were favorable for cephalopod facies accumulation.

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